National University of Life & Environmental Sciences of Ukraine
Department of Soil Science & Soil Conservation

Petrenko L.R., Berezhniak M.F., Kozak V.M., and Berezhniak E.M.

Soil Science with Elements of Geology

A textbook for the students of Agronomy

Kyiv
2015

УДК 631.4:55:[549+552]
ББК 40.4: 26.3:81.43 Я73
П30

Метою видання підручника «Soil Sciemce with Elements of Geology» англійською мовою є посилення навчання студентів ОКР «Бакалавр» напряму підготовки «Агрономія» предмету «Ґрунтознавство з основами геології», який є важливим для їх майбутньої професії. Підручник відповідає навчальному плану підготовки з грунтознавства та геології в українських аграрних вузах. Особливо рекомендується для студентів, що прагнуть опанувати дисципліну англійською мовою.

A textbook has its aim to teach the students of Agrobiology the elements of Soil Science with Elements of Geology important for their profession. It complies with the syllabi in Soil Science and Geology in Ukrainian universities. It is especially recommended for the students whose native language is not English and who wish to learn the course in English.

Рецензенти: декан агробіологічного факультету, д-р с.-г. наук, проф. **В.О. Забалуєв** (Національний університет біоресурсів і природокористування України);
головний редактор наукового журналу НААН України «Вісник аграрної науки», д-р с.-г. наук, проф. **В.А Величко** (Національна академія аграрних наук України);
заступник директора з наукової роботи, д-р с.-г. наук **М.А. Ткаченко** (Національний науковий центр «Інститут землеробства НААН)

Ґрунтознавство з основами геології

Підручник
для студентів ОКР «Бакалавр» напряму «Агрономія»

Автори: Петренко Леонід Романович,
Бережняк Михайло Федорович
Козак Володимир Миколайович
Бережняк Євгеній Михайлович

Contents

PREFICE

This book is an outgrowth of our experiences in teaching the course of Soil Science with Elements of Geology at the Department of Agrobiology of the National University of Bioresources and Life Sciences of Ukraine. The book is recommended for the undergraduate students of Agrobiology (Agronomy) specialized in Crop Production Systems and Agrochemistry and Soil Science. It corresponds to the Typical Program of Study. Our aim with this text is to help preparing the Bachelors of Science in Agrobiology who take some courses of study in English. It also helps to correspond with each other the terminological systems of Geology and especially Soil Science used in Ukraine and elsewhere in the world and thereby prepares the students to understand the original English language courses.

The concepts of soil fertility, productivity, proper use and conservation will be stressed in a series of chapters. Fertilizer use and availability of nutrients are not given much attention, as in Ukraine this subject is taught in the special courses of Agricultural Chemistry and Fertilizing systems.

The first part of the manual is more or less like those taught in the other countries of the world, whereas the second part, especially in the sphere of soil genesis, classification, taxonomy, geography use and conservation remains “more Ukrainian”. It is justified by the circumstance that the book is specially prepared for the Ukrainian. English – language students of Agronomy*. We hope that the readers will find it helpful and not too dull.

L.R. Petrenko compiled the following chapters and their parts: 2,4,5,6,10,22,23,24 and 26.

L.R. Petrenko and V.M. Kozak compiled the following chapters and their parts: 1,3,7,8,9;

L.R. Petrenko and M.F. Berezhniak compiled the following chapters and their parts: 11,12,13,17,18,25, 30;

L.R. Petrenko and E.M. Berezhniak: 14,15,16,19,21;

L.R. Petrenko, M.F. Berezhniak and E.M. Berezhniak: 28 and 31.

Part One.

Geology

INTRODUCTION

Geology (from Greek “Ge”= earth and “logos” = study) is one of the most important sciences about the Earth. We, the humans, share a unique habitat – the earth – with all the other living things. According to our current knowledge, no other planet in our solar system sustains life as we know it. Geology studies composition, structure, and history of Earth, as well as the processes occurring within it and on its surface. Geology is not the only science of Earth. It is connected with numerous natural and historical sciences. Present-day geology uses the methods of mathematics, physics, chemistry, biology, and geography.

A considerable progress in geology’s development started with the appearance of borderline sciences like geophysics, geochemistry, biochemistry, crystallography, paleogeography, etc., which obtain different data on composition, condition, and properties of materials of the earth crust and the geospheres below it. Of primary importance in this respect are geography, oceanography and so on. They allow to understand geologic processes occurring on the earth’s surface, including land forms (relief) formation and the laws governing its development. Geology and geography are mutually “profitable” in the study of land forms. They give birth to geomorphology as an independent science.

Present-day soil science (pedology) was given an impact for its development by geology. Today it is methodologically connected with this science. Thus V. V. Dokuchaev, giving one of his first definitions to the soil, stated that it is “the daylight or external horizons of geologic rocks considerably changed by the combined influence of water, air, and the organisms, alive and dead.” According to V. V. Dokuchaev, a parent material of any soil is one of five natural soil-forming factors, and that it is from parent material that the soil inherits a set of properties so important for the formation of soil productivity, including soil texture (particle size distribution), mineralogical, and chemical composition. Such divisions of geology as petrology (science of the rocks), mineralogy (study of the minerals forming and composing the rocks), and crystallography of mineral solids, give soil science its methods in the study of soil mineralogical composition as well as the laws of soil formation and functioning. The behaviors of water in the soil and soil water regimes cannot be understood without hydrogeology. Soil genesis and evolution base themselves on dynamic geology, particularly on tectonics, seismology, and the study of the volcanoes. Geomorphology helps to understand and evaluate the role of topography (land forms) in soil formation and geography of soils. According to V.I. Vernadsky, the soil on Earth occupies a special intermediate position between the living nature (the organisms) and a non-living one (rocks, minerals, magma, etc.). So the soil is a specific biomineral (living-non-living) body of nature. Space voyages have given us new views of Earth. We are now able to see more details of the earth's surface features. To better understand the earth, we need to observe its features and materials even more closely. The present-day model of the earth's interior allows us to identify the inner core at the center of the earth, the outer core, the mantle and the crust. The outer layer of the earth, or the crust, and the solid layer of the upper mantle, underlying it (the lithosphere, from Greek = stone, rock), make up the direct and special object of geological studies, the subject-matter of the science of geology.

Geology studies the lithosphere's composition in the form of geologic rocks, their minerals and chemical elements composing them. Some rocks from the hardening and crystallization of magma (the molten solid formed near the bottom of the crust or within the upper mantle). Such rocks are known as igneous rocks.

Some rocks form at the surface from exposed igneous and other rocks, undergoing weathering, transformation, and sedimentation. They are known as sedimentary rocks. Still other rocks form under conditions of high heat and pressure, deeply buried, from the former igneous or sedimentary rocks. They are known as metamorphic (changed) rocks.

The lithosphere's material composition is studied by the geological sciences partly united under the name of the sciences of geochemical cycle. They include Petrography (from Greek "petras" = rock, stone and "grapho" = to write or describe) or Petrology, which is the science of igneous and metamorphic rocks, their composition, structure, conditions of formation and the extent of changes undergone under the impact of various factors, as well as their distribution within the earth's crust. Lithology is the science which studies sedimentary rocks. Mineralogy is the study of the minerals composing the rocks. Crystallography and Crystalochemistry study mineral crystals. Historical geology is the study of the origin and evolution of the earth and its inhabitants. In working out the geologic history of an area, the geologist utilizes Stratigraphy, which is concerned with the origin, composition, sequence, and correlation of the rock strata, Paleonthology (the study of ancient organisms as revealed by their fossils), and Paleogeography, which deals with the studies of the geographic conditions of past times. Quaternary geology studies the deposits of the latest of geologic periods which goes on today. Hydrogeology is the science of the groundwater system, which is an important part of the water cycle and the use of water resources, including water conservation.

Geochemistry is a generalizing branch of science on the chemical composition of the lithosphere. It also studies the history of chemical elements and the laws governing their distribution and migration in the interior and the surface of the earth.

In order to understand the earth, we need to observe it in many different ways. Our present knowledge has come from countless observations and investigations over many years. Earth has been observed from the atomic level to its cosmic level. The process by which scientists use satellites to observe the earth is called remote sensing. Scientists use satellite photographs made from the images to study earth resources. Crop yields are determined by studying soil, plant conditions, and climates depicted on each photograph. Geologists can identify mineral deposits by observing land structures. Water engineers and environmentalists directly observe river patterns or amounts of pollution in the bays and lakes.

A future agronomist should understand the causes and motive forces of the processes of wind and water erosion of soils, as well as to know which rocks and minerals are utilized as raw materials for the production of commercial fertilizers and soil amendments.

Questions and Assignments:

1. Why is geology considered one of the most important sciences of the Earth?
2. Name and define the main divisions of geology.
3. Which sciences and why comprise the geochemical cycle?
4. Speak on the importance of geological information for the students of agrobiology and environmental studies.

Chapter 1

General Views of Earth

1.1. Earth in the Solar System and the Universe

Earth as a celestial body is not the subject of geology. But to understand geology we must have the general notions on the Earth's position in the solar system and the Universe, as the environment, which surrounds our planet, affects the processes within the earth and on its surfaces. Our galaxy, the Milky Way, is a bright band of light which stretches from horizon to horizon. Early Greek astronomers guessed that the Milky Way was made of many stars. In 1610 Galileo found they were right when he viewed the area with his telescope.

The Milky Way is a spiral galaxy. Its diameter is about 80,000 light-years. It contains about 100 to 200 billion stars. Almost all of its stars are at least 100 light-years from Earth. Alpha Centauri is our nearest star. Yet, light from Alpha Centauri left there 4.3 years before we see it. In our galaxy, there are about 100 clusters of stars. These clusters contain thousands of stars. The center of the galaxy is hidden by dark clouds of gas. The sun is near a spiral arm about 30,000 light-years from the center of the Milky Way. The sun carries the whole solar system with it as it orbits the center of the galaxy at about 200-300 kilometers per second. The orbit takes about 250 million years to complete. The sun is a main sequence star. It came into being about 4.6 billion years ago. Mercury, Venus, Earth, and Mars are the inner planets. These planets are close to the sun. All of them are solid bodies in contrast to the outer planets, most of which are gaseous bodies. Mercury, Venus, and Earth are from the sun at the distance of 57.94, 108.26, and 149.509 mln kilometers respectively. After Mars, there come asteroids, Jupiter, Saturn, Uranus, Neptune, and Plato. Comets and meteoroids are other bodies of the solar system. Some solar system data are given in the following table (Table 1).

Earth has a diameter of 12756 km. Moon is its only satellite. Earth and its moon travel around the sun as a unit. The distance between them is of 384400 km. The pull of Moon's gravity is only one-sixth of the Earth's gravity. Since the rotation and revolution of the Moon take about the same amount of time, a casual observer on Earth always sees the same side of the Moon. The Moon's diameter is 3476 kilometers. The Moon's density is 3.3 g/cm^3. Because of its low density, the Moon does not have a large enough gravity field to hold an atmosphere.

The sun contains 99.86 percent of the mass of the solar system. Most of the mass of the Sun is due to the gases, hydrogen and helium. Gases at the sun's equator move faster than those at the poles. The sun's diameter is 1341 thousand kilometers which is about 109 Earth diameters. The temperature of its surface is of 6000°C and that of its interior, according to astrophysicists, reaches 20,000,000°C. The sun has a central gaseous core and several other layers (Fig. 1).

Table 1. Solar System Data (Focus on Earth Science, 1991)

Characteristics	Mercury	Venus	Earth	Moon	Mars	Jupiter	Saturn	Uranus	Neptune	Pluto	Sun
Mean distance from Sun 10^6 km	58	108	150	–	228	779	1428	2869	4496	5906	–
Length of year (Earth days)	87.96	224.7	365.24	–	686.98	4332.6	10759.2	30685	60190	90700	–
Period of rotation (days)	58.65	243	1.0	27.3	1.02	0.41	0.43	0.45	0.6	6.4	25.4
Mean density g/cm^3	5.5	5.2	5.52	3.34	3.9	1.33	0.7	1.71	1.77	1.5	1.4
Diameter at equator (km)	4880	12100	12756	3476	6786	142200	119300	47100	48400	2414	1392000
Relative mass (Earth = 1)	0.06	0.82	1.0	0.12	0.11	318	95	14.6	17.3	0.002	–
Surface temperature (°C)	350 to -170	500	–	–	-30 to -85	-146	-180	-190	-220	-230	–
Number of Satellites	0	0	1	–	2	13 or 14	10 or 11	5	2	1	–
Atmospheric gases	Hl, O_2, H_2	CO_2, HCl, CO, H_2O	N_2, H_2O, O_2, Ar, CO_2	–	CO_2, N_2, O_2, Co, Ar, H_2O	H_2, Hl, CH_4, NH_5	H_2, Hl, CH_4, NH_3	H_2, Hl, CH_4	H_2, Hl, CH_4	Frozen gases	–

The sun's gases are in constant motion which makes it impossible to tell exactly where boundaries are between different layers. In the core, hydrogen gas is changed into helium gas. At temperatures of about 15 million °C, protons, which normally repel each other, are forced to combine. A stable He nucleus of two protons and two neutrons is formed. During this nuclear reaction, energy, which is the source of the sun's heat and light, is released. Radiation flows outward from the sun's interior to the surface. From the surface, radiation escapes into space. The photosphere is the surface of the sun. This layer emits visible radiation. The chromosphere is a bright red halo which extends about 6000 km beyond the photosphere. The red color is due to the hydrogen gas presence. Prominences of gas shoot outward from the chromospheres. Some fall back into the sun. Others escape into space. Beyond the chromospheres is the thinner, transparent zone of the corona. This zone of very hot gas is visible only during a total eclipse of the sun. Sunspots are cool, dark areas on the sun's surface. Temperatures in sunspots are about 4000°C. Solar flares are sudden increases in brightness near sunspots groups. They disturb radio reception and affect Earth's magnetic field. Gases in the upper atmosphere excited by solar flares, radiate lights known as the aurora borealis. The solar wind is made up of ions moving outward from the sun. The velocity of the solar wind ranges from 250 to 800 km per second. The solar wind is especially strong after solar flare activity. Ions are deflected toward Earth's poles by the magnetic field. The solar wind interferes with radio waves in the upper atmosphere. The galaxy of the Milky Way moves towards the stellar system of the Unicorn at a velocity of 210 km per second. All the galaxies are combined into Megagalaxies and the Universe is endless.

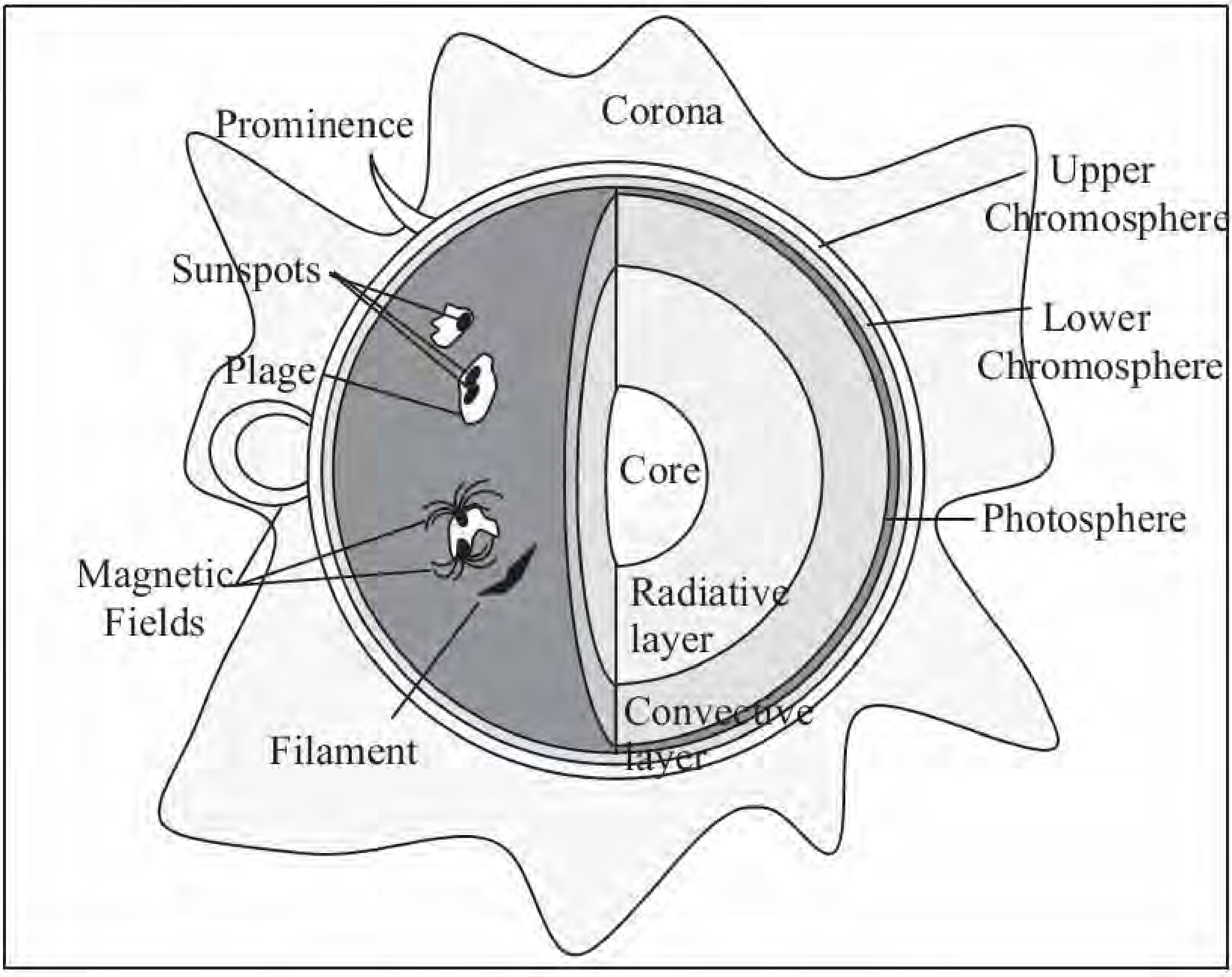


Fig. 1. Structure of the sun (by Riza Arievanda).

1.2. Origin of the Solar System and Earth

Earth is the third planet outward from the sun. Its origin has been debated for a long time. Many questions still are unanswered. Probably the most widely accepted theory for the origin of the solar system is the dust cloud theory. According to this idea, a dust cloud began to rotate. Movement grew faster and faster until the dust and gas flattened into a disk-shaped cloud which separated into eddies. These eddies were similar to the ones seen in swiftly moving water. At the center of the cloud, the largest eddy formed the sun. In the smaller eddies, grains of matter grew larger by collision. The grains stuck together and grew larger as they gathered up more material. When the mass had swept up most of the material in the eddy, a planet was formed.

Questions still rise about the moon origin. The idea of Moon capture by the earth is the most widely accepted idea at present. Probably many moon-sized bodies formed in the original dust and gas cloud. These bodies combined with the planet or moved into space depending on how close they were to the earth when they formed. The moon must have been a rigid body when it joined the earth-moon system. Only a rigid body could maintain its spherical shape despite the earth's gravity field.

1.3. Shape, Dimensions, and Structure of Earth

An oblate spheroid called geoid is the best description of the earth's shape. Recent data from satellites suggest that the earth's southern hemisphere is a little larger than the northern hemisphere. The earth bulges at the equator. The oceanic floor may have trenches (gashes cut into the abyssal plain). Many ocean canyons are longer and deeper than anything on the continents. The Grand Canyon is about 1.6 kilometers deep. The Marianas Trench is at least 11 kilometers deep. This trench is near the Marianas Islands in the Western Pacific Ocean. High mountain ridges also make the earth's surface rough. Mount Chomolungma (Jomolungma) in the Himalayas is almost 9000 meters high (8848 m). It is the loftiest peak of the entire world. Below are given the basic data characterizing Earth:

- Equatorial radius (km) 6378.2;
- Polar radius (km) 6356.9;
- Surface area (mln km^3) $1.08*10^{12}$;
- Mass of living matter in the biosphere (tons) $5.0*10^{12}$;
- Mean density (g/cm^3) 5.52;
- Mean density of surface rocks (g/cm^3) 2.7-2.8.

The earth's crust is a rock wall of the globe. It is studied by geology. Hydrosphere is studied by Hydrology and Oceanology and the atmosphere – by Meteorology and Climatology.

1.4. Earth's Spheres

The body of Earth has a concentric structure. The core is in the center of it. The geospheres concentrically surround it. Earthquake studies give us the only clues to the composition or makeup of the inner layers of the earth. These are the layers of different density. The geospheres are internal and external. Atmosphere, hydrosphere, and biosphere belong to the external or outer geospheres. The earth's crust, mantle and core belong to internal geospheres. The rotation of the earth on its axis is the main cause for its differentiation into the geospheres. The density of the spheres increases from the surface to the center of the globe.

1.4.1. Outer Geoshperes

The earth's blanket of the atmosphere is a mixture of many gases, mainly nitrogen 78%, and oxygen, 21%. Other gases such as carbon dioxide and water vapor vary in amount from place to place (Table 2).

Table 2. Gases of Earth's Atmosphere

Gas	Chemical symbol	Percent by volume
Nitrogen	N_2	78.09
Oxygen	O_2	20.95
Argon	Ar	0.93
Carbon dioxide	CO_2	0.03
Water vapor	H_2O	0 to 4.0
Neon	Ne	trace
Helium	He	trace
Methane	CH_4	trace
Krypton	Kr	trace
Xenon	Xe	trace
Hydrogen	H_2	trace
Ozone	O_3	trace

It is believed that in the last decade or so the content of CO_2 has increased from 0.03 to 0.045%. The gases in the air are involved into cycles. They are absorbed by the plants and consumed by the animals, and then again come back to air, water or rocks. Water vapor in the air is a source of precipitation and it also increases the air's density. The atmosphere partly absorbs and disperses the solar radiation. Of the total amount of solar radiation incoming to the earth, about 31% is reflected back into outer space and 69% are utilized for the warming up of the atmosphere and the surface of the continents and oceans. Atmospheric pressure is of great importance for the movements of air masses. Air masses distribute the evaporated water rather unevenly over the globe. The weather is dependent on them. Water is essential for life processes. Water vapor, like carbon dioxide, helps prevent heat loss from the earth.

The atmosphere is divided into four layers based on temperature. The troposphere is the layer nearest Earth. It contains 75 percent of the gases of the atmosphere as well as dust and water vapors. This layer is the zone where weather occurs. Temperatures decrease with increasing height in the troposphere. Near the top of the troposphere, about 11 kilometers from Earth, a thin layer called the tropopause acts as a ceiling to the weather zone. This layer contains strong winds called the jet streams. Above the tropopause, lies the stratosphere. This layer extends upward to about 50 km from the earth. In the lower part of the zone, temperatures are a constant -50°C. However, at about 50 km temperatures rise to about 0°C. The stratosphere contains layer of ozone. Ozone keeps most of the ultraviolet radiation from the sun from reaching Earth's surface.

The mesosphere, the next layer of atmosphere, is the coldest zone. This layer extends from about 50 km upward to about 85 km. Temperatures decrease to near -100°C at the top of the mesosphere.

The next layer of atmosphere is the thermosphere which includes the ionosphere and exosphere. The thermosphere extends upward to 500 or 750 km. Temperatures in this zone increase rapidly because of the absorption of solar energy. The ionosphere is electrically charged. Ions form three layers that reflect radiowaves back to Earth. The radio waves are reflected at an angle and return to the earth far from their starting point. In the lower part of the exosphere, the atmosphere becomes extremely thin. Some gases escape into space. The atmosphere extends possibly as far as 80000 km beyond the earth.

But to the geologists, troposphere is the most important layer whose thermal regime is mainly caused by the heat obtained from the earth's surface. Every 100 meters of elevation decrease air temperature by 0.5-0.6 degrees, so that at height of 10-12 km it is about -55°C.

Hydrosphere includes the ocean, a continuous body of salt water that covers a little over 70 percent of the earth's surface. Large as the ocean is, it is only part of the hydrosphere, or Earth's water portion. Water in lakes and rivers and ice frozen in glaciers are all part of the earth's water supply. Ground water is closely connected with the hydrosphere. According to V. I. Vernadsky, oceanic water occupies a volume of 1370 mln km^3. Inland water is of 4 mln km^3, continental glacier water of 16-20 mln km^3, and ground water – of 400 mln km^3. Total volume of natural water on Earth is about 1.8 billion km^3 Mean depth of the hydrosphere is of 3.75 km, the greatest being of 11 km^3 (the already mentioned Marians Trench). Chemical elements dominating in the hydrosphere are oxygen and hydrogen (96.69%), and sodium and chlorine (3%). The hydrosphere interacts with the other geospheres, so all elements of the periodic system are encountered in it. Composition of ocean water is not constant. Some substances are carried to the ocean by rivers that flow over weathered rocks. Many elements are in solution in the water vapor given off by volcanoes. Chlorine gas is one material that is added by volcanic action. Sodium is a product of weathering and probably is carried to the ocean by surface waters. Other substances dissolved in seawater include calcium, magnesium and potassium. When ocean water evaporates, it leaves the dissolved salts behind. Salt, magnesium, and bromine are three products extracted today from seawater.

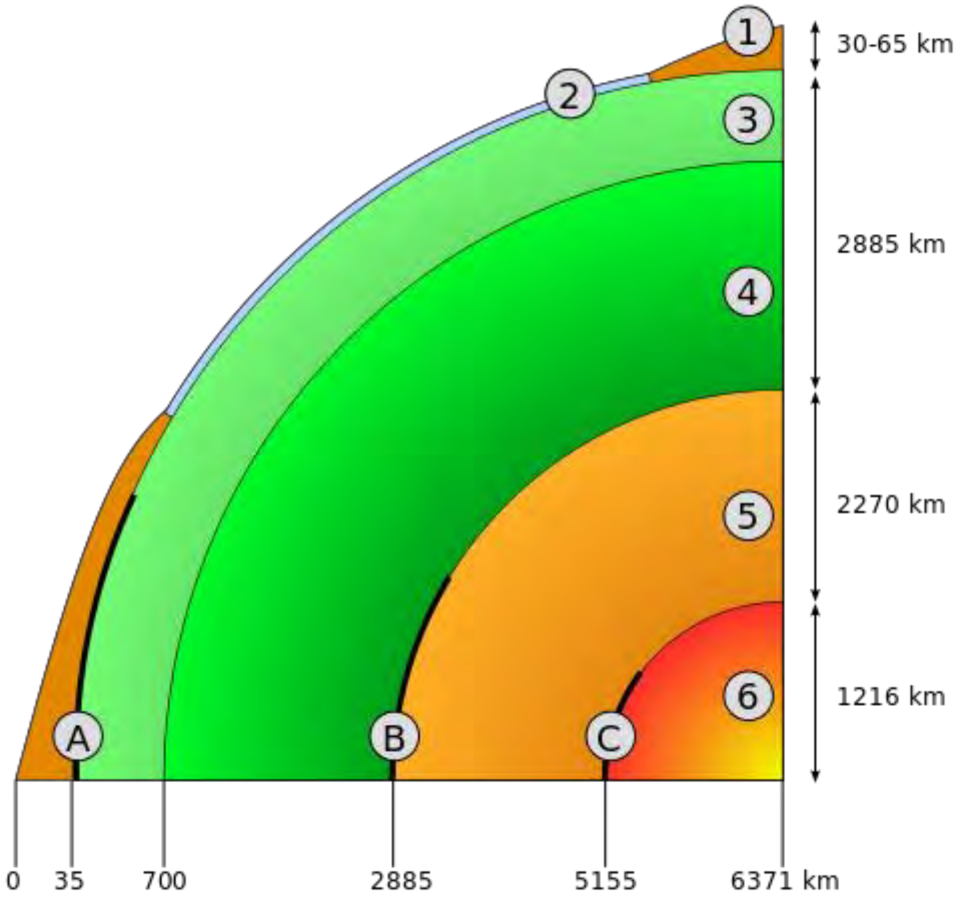
The biosphere is the zone of the organisms' active life. It occupies the lower atmosphere (12-16 km), hydrosphere, and the upper layers of the earth's crust (about 3-4 km). Its total thickness may reach some 25-30 km. Biosphere takes part in the processes of rock formation and destruction. According to V.I. Vernadsky, there is no chemical force on Earth more constantly acting and more powerful by the end product of its action, as the living organisms.

According to some calculations, the living matter of Earth is about 10^{12} metric tons in mass which is no more than 1/100,000 part of a percent of the mass of the earth's crust. Ecology is the study of the relationship between an organism (living thing) and its environment. No organism is completely self-sufficient. The biosphere is made up of many ecosystems. An ecosystem can be as small as the surface of a leaf or as large as the entire ocean.

Today's Earth is a habitat for over 500,000 plants and over 1mln animal species. All geospheres are densely populated by the microorganisms. They participate in the cycles of life (small biological cycle of matter and energy on Earth). Soil formation is a part of this kind of activity. Soil cover of the earth is often referred to as pedosphere. The soils provide physical support for plants and create the water-and-air-filled pores that make plant growth possible and contain the elements of plant nutrition.

1.4.2. Internal Geospheres

The internal structure of the earth can be presented by the following scheme (Fig. 2).



1. Continental crust; 2. Oceanic crust; 3. Upper mantle; 4. Lower mantle; 5. Outer core; 6. Inner core; A: Mohorovicic discontinuity; B: Gutenberg Discontinuity; C: Lehmann–Bullen discontinuity.

Fig. 2. Internal Structure of Earth (by Dake, From Wikipedia).

The Earth’s crust is the solid layer of Earth. Compared with other layers, it is the most nonuniform in composition and very dynamic. The tectonic theory proves that the earth’s crust is made up of the floating plates, which “float” or move very slowly (a few mm per year) on the molten material of the mantle.

The crust experiences a lot of stress caused by forces within the Earth itself, as the crust is the surface, or outermost, layer of the Earth. There are two major sections (types) of the crust. One section is called continental crust. Continental crust makes up the Earth’s landmasses, such as the Eurasian or the North American continent. In most places, continental crust is about 32 km thick. But under tall mountains, it can be up to 70 km thick. The other section of the crust is called oceanic crust. It is found under the ocean floor. It is thinner than continental crust, and may be about 8 km thick.

Continental crust usually consists of three layers: sedimentary, granitic, and basaltic (Fig. 3). Oceanic crust usually consists of two layers: sedimentary and basaltic.

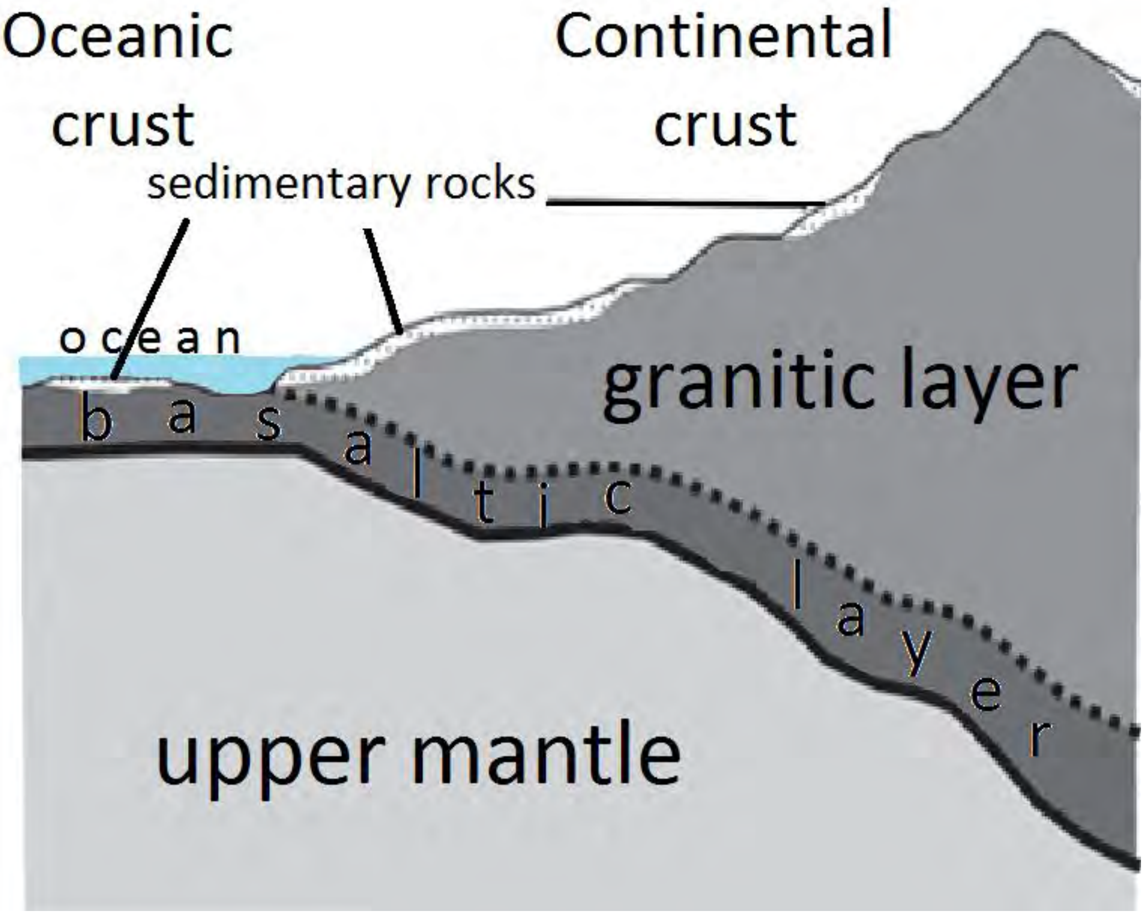


Fig. 3. Structure of the Earth’s Crust.

Sedimentary layer is composed by soft clastic or other sedimentary rocks. The majority of such rocks have layered structure. Their density varies from 1.00 to 2.65 g/cm^3. The thickness of sedimentary layer may vary from a few meters to 10-15 km. There are places on Earth where sedimentary layer is absent.

Granitic layer is composed mainly by the igneous and metamorphic rocks with the dominant content of aluminum and silicon in them. Mean content of silica (SiO_2) in the rocks over 60% and they are, therefore, acid rocks. Rock density is within 2.65-2.80 g/cm^3. The layer’s thickness is not the same everywhere, the greatest (60-70 km) being encountered under mountain ranges.

Basaltic layer directly underlies a granitic one. Its thickness may vary from 5 to 30 km. By chemical composition, its rocks are basalt-like, basic ones with lower silica content than that of granites. The layer's density may reach 3.32 g/cm^3. The lower boundary of this layer is known as Mohorovicic discontinuity, or the Moho. This sharp boundary, first identified in 1909 by A. Mohorovicic, a Yugoslav seismologist, lies 30 to 50 km beneath the surface.

Beneath the Earth's crust is a layer called the mantle. It extends from the bottom of the crust downward make it longer about 2900 km to the Earth's core. The mantle is made of rock that flows slowly-sort of like molasses or hot, thick tar. Because the mantle is much denser than the crust, the solid rocky crust floats on the mantle. It is distinctly divided into three layers: upper (B), intermediate (C) and lower (D), the last being at depth of 900-2900 km. The upper mantle is composed primarily of oxygen, silicon, and probably magnesium. The lower mantle is composed primarily of oxygen, silicon, magnesium, and nickel. The specific gravity of the rocks in the mantle ranges from 3.5 (in the upper part) to as much as 8.0 at the bottom. Some scientists give evidence of the D-layer density to be within 5.7 to 9.4 g/cm^3. The core of the earth has a density approaching that of iron at appropriate pressure. So some think that the core contains iron and nickel and is even endowed with magnetic properties. The super-high pressure within the ore impedes the melting of its material and it possesses the properties of heavy metals. But all these are the hypotheses. The core is thought to be very hot, dense, and under tremendous pressure. The outer core is believed to be fluid. The materials composing this zone have specific gravities of 12.0 or more. The inner core is believed to be solid. The rocks are presumed to be quite heavy. Some may have a specific gravity of more than 17 (17.3-17.9 g/cm^3). All the divisions of the earth were identified by the velocity of seismic waves. They move through the Earth at different speeds, depending on the density of the material through which they are moving. As they move deeper into the Earth, where material is more dense they speed up.

Questions & Assignments

1. Shortly characterize the position of Earth and Solar system in the Universe.
2. Give a short characteristic of the sun.
3. Speak on the shape and dimensions of the Earth.
4. Shortly characterize the external geospheres.
5. Tell whatever you know about the earth's crust
6. What do you know about the internal structure of the Earth?
7. What is the biosphere and what function does it fulfill on Earth?
8. Speak on the continental and oceanic sections (types) of the earth's crust.

Chapter 2

Geological Processes

2.1. Endogenic and Exogenic Processes

What we in our country call endogenic and exogenic processes are known in the English-language countries as the processes of internal and external dynamics, respectively. Along the entire history of the Earth it was subjected to dynamic changes. The composition of the earth's crust, shape and even the number of continents, land forms, etc., were in a state of continuous change.

The processes of internal and external dynamics take place continuously and at the same time, creating all the multiformity and variety of the structure and composition of the earth's crust. The processes of internal dynamics include the earth's crust dynamics, plate tectonics, magmatic processes, metamorphic processes, volcanoes and earthquakes. The external dynamics processes include weathering, mass wasting, and the geologic activities of the rivers, wind, ground water, glaciers, oceans, bogs, lakes, etc. Fault and fold deformations in the earth's crust are connected with origin of mountains (orogenesis). Exogenic processes occur in the surface layers of the crust and on its surface under the influence of the solar energy. Gravity forces also contribute to the character and intensity of these processes. The heat that evolves during the processes of radioactive decay complimented by the energy of gravity makes the motive force of endogenic geological processes.

Both exogenic and endogenic processes occur with a differing rate which is generally relatively low. That is why the action of these processes can make itself felt in the course of many hundreds, thousands, and even millions of years. But there are such events as the earthquakes, volcanic eruptions, landslips, mudflows, etc. They occur catastrophically rapidly.

The interaction of internal and external processes in the earth's crust and on the surface of the Earth forms the deposits of various useful materials, belonging to nonrenewable resources. Coal, oil, gas, salts, etc. form their deposits as a result of exogenic processes. Endogenic processes connected to the intrusions of magma and its differentiation into various materials, as well as metamorphic processes, usually lead to the formation of metal or other ores. Internal dynamics processes, certainly, play a greater role in the formation of the earth's crust per se and in the manifestation of continental drift and other phenomena of plate tectonics.

Many scientists pay attention to the fact that endogenic and exogenic processes act upon the earth's crust in opposite directions. The former create the landforms on the surface of the earth, whereas the latter tend to smooth them out. Soil erosion and terrestrial denudation are sometimes viewed by the scientists as the great leveler of the land. We do not only study the Earth as we see it today, but also its history as it has evolved to its present condition. We must look at processes and structures that occur today, and interpret what must have happened in the past. Then we shall have a better understanding of things happening around us.

2.2. Internal Dynamics Processes

What causes the land to change? How do the mountains and the valleys form? Throughout the Earth's long history, its surface has been lifted up, pushed down, and broken by forces beneath the surface. Although the movements of the Earth's surface are usually too small and too slow to be directly observed, they are constantly changing the appearance of the Earth.

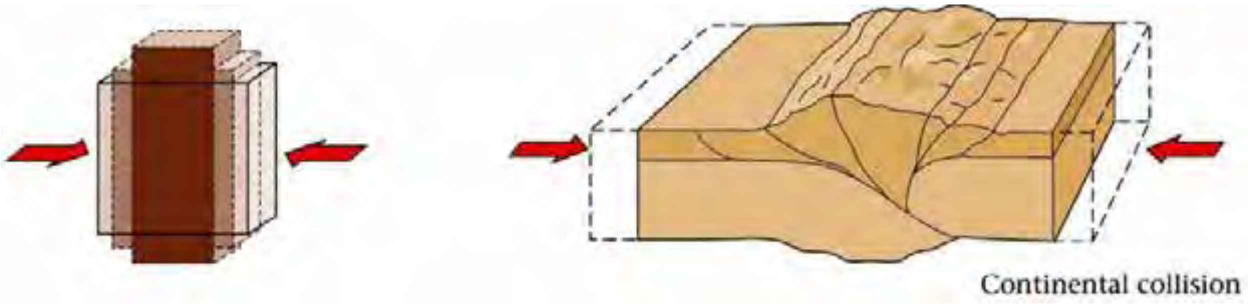
These movements may be slow (epeirogenic) and violent (earthquakes and volcanoes). Thus the Earth looks different today from the way it did millions of years ago. What were once small hills may now be mountains that stand almost 9 kilometers above sea level!

The Earth also experiences stress. It is caused by forces within the Earth itself. These forces push and pull on the part of the Earth known as the crust. You already know that the crust is the surface, or outermost, layer of the Earth. There are two major sections of the crust. One section is called continental crust. Continental crust makes up the Earth's landmasses, such as the North American or Eurasian Continents. In most places, continental crust is about 32 kilometers thick. But under tall mountains, it can be up to 70 kilometers thick.

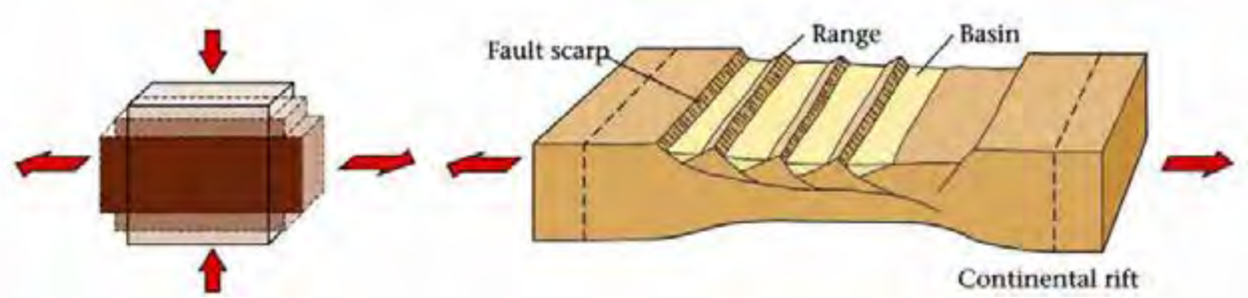
The other section of the crust is called oceanic crust. Oceanic crust is found under the ocean floor. It is thinner than continental crust. For example, it may be about 8 kilometers thick.

As the rocks of the crust undergo stress, they slowly change shape and volume. They also move up or down or sideways. Their movement causes the rocks to break, tilt, and fold. The breaking, tilting, and folding of rocks is called deformation. The three types of deformation are compression, tension, and shearing, which are illustrated by the following figure:

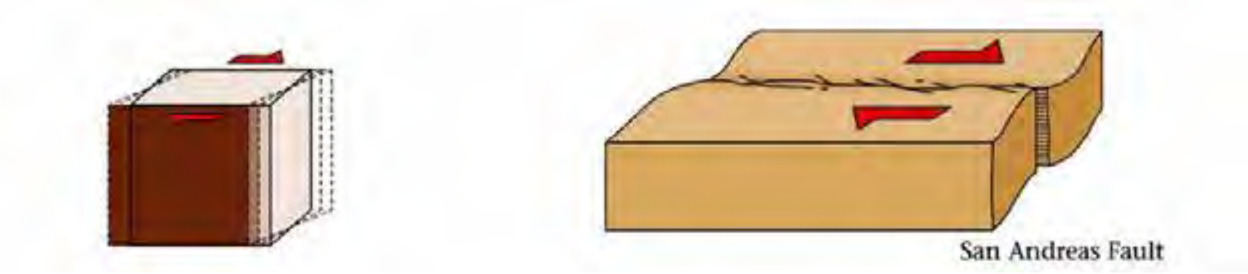
Compression squeezes the rocks of the crust. The particles in the crustal rocks move closer together, making the rocks denser and smaller in volume. Compression is acting like a trashcompacter, squeezing a large amount of matter into a smaller amount of space. Tension pulls on the rocks of the crust causing them to stretch out over a larger area. A rock becomes thinner in the middle than at the ends. As the volume of the rock increases, its density decreases. This causes the rocks to twist or tear apart. During shearing, the rocks simply bend or break apart.



(a) Compression



(b) Tension



(c) Shearing

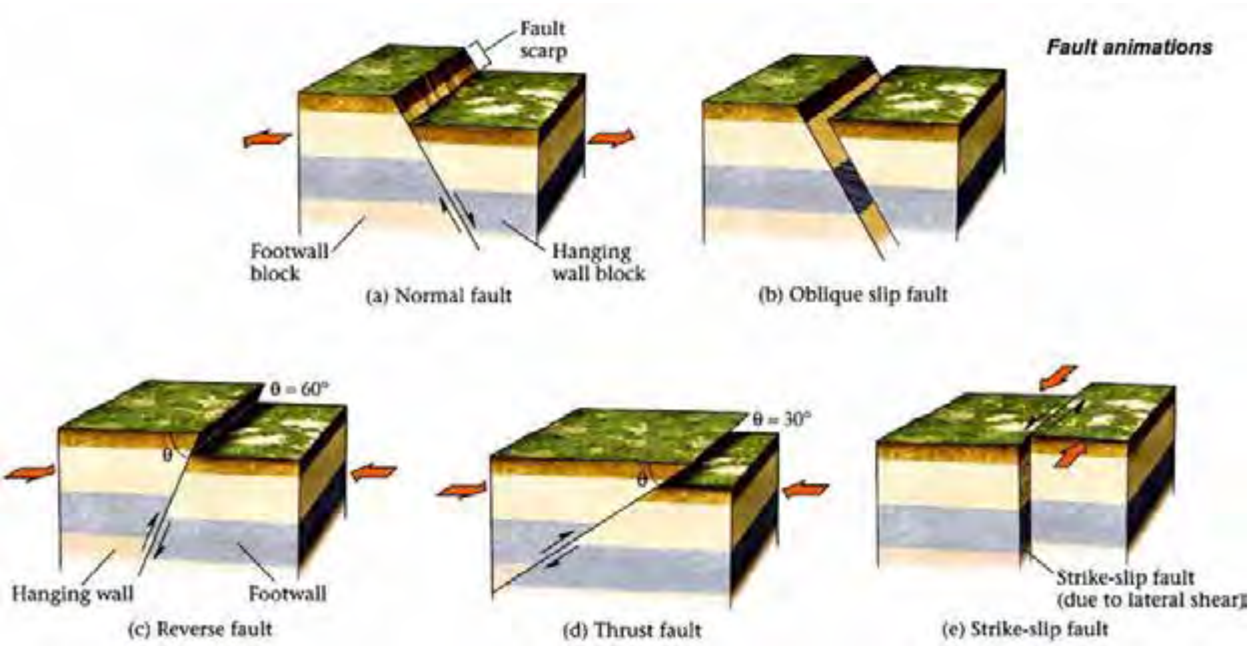
from "Crustal Deformation and Mountain Building", by Mt. Cook, New Zealand

Fig. 4. Crust Deformations by Different Forms of Stress.

Compression, tension, and shearing can change a rock's volume, its shape, or both. These stresses can also cause the rocks to fracture, or crack. If the rocks fracture along numerous flat surfaces which show no displacement, the rocks are called joints. Joints are generally parallel to one another. Some rocks have joints that form in more than one direction. Such rocks may break into blocks.

2.2.1. Tectonic Movements of the Earth's Crust.

Faulting. Stress sometimes causes rocks to break. A break or crack along which rocks move is called a fault. The rocks on one side of the fault slide past the rocks on the other side of the fault. Movements along a fault can be up, down, or sideways. Earthquakes often occur along faults in the Earth's crust.



from “Crustal Deformation and Mountain Building”, by Mt. Cook, New Zealand

Fig. 5. Types of Fault.

The block of rock above the fault is called the hanging wall. The block below the fault is called the foot wall. Stress can cause either the hanging wall or the foot wall to move up or down along a fault. If tension is acting on a fault, the hanging wall will move down relative to the foot wall. If this occurs, the fault between the two blocks is called a normal fault. If compression is acting on a fault, the hanging wall will move up relative to the foot wall. This type of fault is called a reverse fault. A special type of reverse fault is a thrust fault, when compression causes the hanging wall to slide over the foot wall. Thrust faults usually carry rocks many kilometers from their original position. A thrust fault pushes older rocks on top of younger rocks.

Shearing will cause the blocks of rock to slide horizontally past each other. One block moves to the left or right in relation to the other block. The fault along which the blocks move horizontally past each other is called a lateral fault:

The lateral fault is also known as a strike-slip fault, as the blocks of rock move horizontally past each other.

from “Crustal Deformation and Mountain Building”, by Mt. Cook, New Zealand

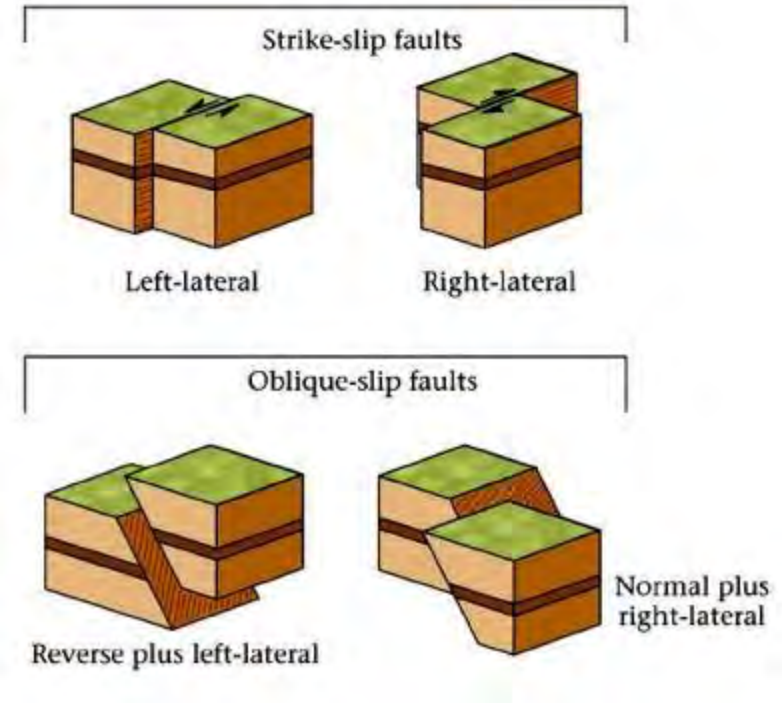
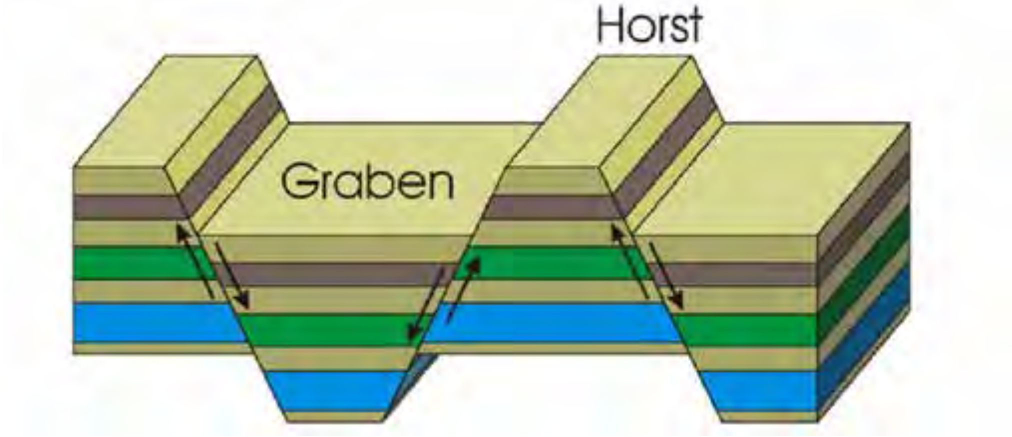


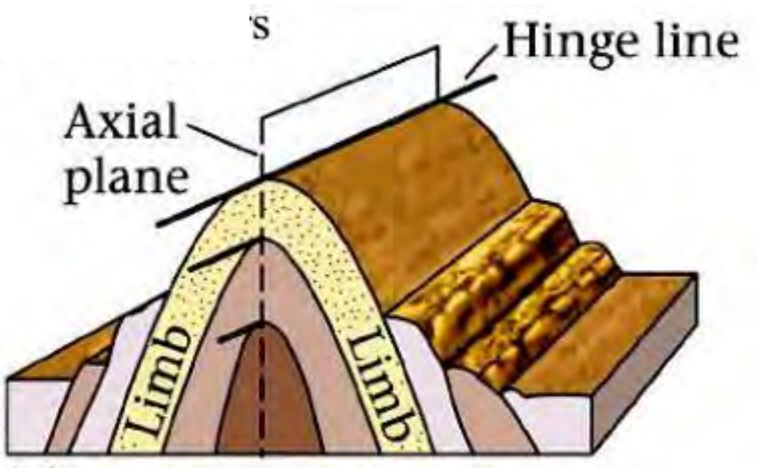
Fig. 6. Lateral Fault.

Faulted Mountains and Valleys. When there are many normal faults in one area, a series of mountains and valleys may form. Mountains formed by blocks of rock uplifted by normal faults are called fault-block mountains. Thus the Cordilleran Mountain region contains many fault-block mountains. Valleys also form when mountains form. Some valleys are formed when the block of land between two normal faults slides downward. Valley created in this way are called rift valleys, like Death Valley in California. It is a long, narrow valley 87 meters below sea level.

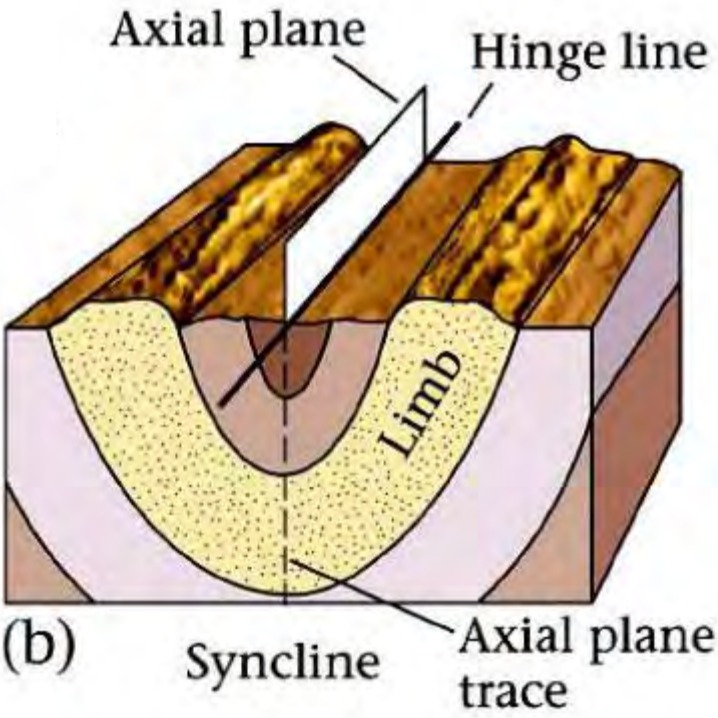
Folding. Sometimes when stress is applied to the rocks of the crust, the rocks bend but do not break. A bend in a rock is called a fold. A rock can fold either upward or downward. An upward fold in a rock is called an anticline. A downward fold in a rock is called a syncline.



Fault Structure (Horst and Graben)



(a) Anticline



(b) Syncline

from “Crustal Deformation and Mountain Building”, by Mt. Cook, New Zealand

Fig. 7. Faulting and Folding.

Folds vary in size. Some are large enough to form mountains. Layered rocks with large folds often have smaller folds within the layers. The Appalachian Mountains are made of many anticlines and synclines. This folded mountain chain extends from Canada to Alabama. Even though an anticline is an upward fold, it is not always higher than the surrounding land. An anticline may be under hills, valleys, or flat areas. An anticline may be hidden by layers of rock that build up in the low-lying areas around it after it forms. Or the stress may not have been great enough to bring the fold to the Earth’s surface.

A plateau [plaʻтɔ] is a large area of flat land that is raised high above sea level. Plateaus are often raised up by the same processes that form mountains. But the rock layers in a plateau remain flat. This is not the case with mountains, in which the rock layers are tilted and broken by faulting or are warped by folding. One way a plateau may be formed is by a slow flat-topped fold. The Appalachian Plateau, which lies just west of the folded Appalachian Mountains, was created millions of years ago by such a fold. This plateau covers much of New York, Pennsylvania, Ohio, Kentucky, West Virginia, and Tennessee.

Plateaus can also be formed by a series of molted rock flows on the surface of the Earth. Molten rock at the surface of the Earth is called lava. Molten rock deep within the Earth is called magma. The lava sometimes fills in valleys and covers hills. The flowing and spreading out of lava is repeated over and over again. The hardened lava sheets pile up and form a raised plateau. Rivers often carve large plateau into many smaller plateaus or cut deep valleys and canyons through plateaus.

Domes. Sometimes magma pushes upward but does not reach the Earth's surface. The stress caused by the magma caused the rock layers above it to fold upward, forming an uplifted area. At some point, the magma cools and forms hardened rock. The uplifted area created by rising magma is called a dome. A dome is a raised area shaped roughly like the top half of a sphere. The outline of a dome is oval or circular. You can think of a dome as rather like a blister on the surface of the Earth.

Magma. Over a long period of time, the uppermost rock layers may be worn away to reveal the dome's core of hardened magma.

from "Crustal Deformation and Mountain Building", by Mt. Cook, New Zealand

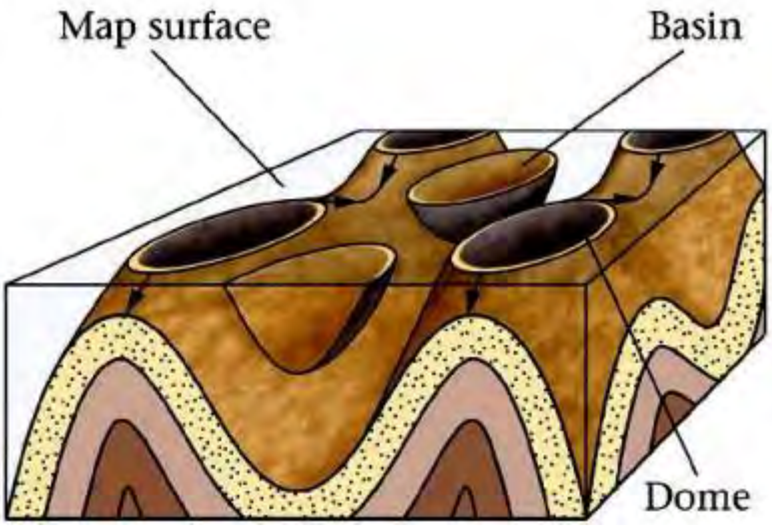


Fig. 8. A Dome.

Domes may form many separate peaks called dome mountains. The Black Hills of South Dakota and Wyoming are dome mountains.

The majority of mountains formed as folded mountains. They include the Himalayas, The Caucasus Mountains, the Andes, and the Pyrenees. Folded mountains were formed in different epochs. The Scandinavian and Scotland mountains are the oldest. The Urals, Altay, Tien-Shan, etc., were formed later on. But the folds can easily be seen in some younger mountains like the Caucasus, the Alps, and the Himalayas.

A number of factors determine if the rocks will fault or fold. One factor is temperature. IF the rocks become extremely hot during compression they are more likely to fold than to fault. Another factor that affects whether rocks will fault or fold is pressure. The greater the pressure applied to the rocks, the more likely they are to fold rather than fault.

Rock type is yet another factor that determines whether rocks will fault or fold. Some types of rocks break easily when stress is applied. Such fragile rocks are said to be brittle. Sandstone is one example of brittle rock. Other rocks, such as rock salt, bend easily under stress. Rocks that bend easily are said to be ductile. Ductile rocks are more likely to fold, whereas brittle rocks are more likely to fault. Still another factor is whether the stress is applied gradually or suddenly.

Deep Fractures in the Earth's Crust. Such fractures could have taken place at the dawn of our planet formation. They may be billions of year old. The fractures may reach the depth of 800 kms and more. Their length may be up to a few thousand kilometers. The entire earth's crust is divided by them into large approximately rectangular blocks. The belts of basic and ultrabasic intrusions are connected with deep fractures and, therefore, they form the fields rich in useful fossil materials. Deep fractures can be identified even within the Ukrainian Crystalline Shield. Deep fractures extend themselves also along the ocean floor. Their origin still remains enigmatic. The evolution of a large amount of heat during radioactive decay processes in the deeper parts of the earth's crust may cause the distension and cracking of the upper parts of it. Be it as it may, but they theory of plate tectonics should be taken into the arsenal of our science too.

In spite of all the evidence from fossils and rocks, many scientists still refuse to accept the theory of continental drift. During the 1950's and 1960's, new techniques and instruments enabled scientists to make better observations of the ocean floor. The plates of the earth's crust meet at plate boundaries or margins. There are three types of plate margin:

1) Diverging plates causes constructive plate margins: e.g. Mid-Atlantic Ridge. Plates move apart. Magma oozes and forms new plate;

2) Converging plates cause destructive plates margins: e.g. West coast of South America. Two plates move toward each other. Crust is destroyed. Fold mountains, earthquakes and volcanoes are common;

3) Transform plates cause conservative plate margins: e.g. the West coast of North America. Plates move sideways against each other. Material is not gained or lost. Volcanoes are rare but earthquakes are common. Friction between the two plates causes earthquakes.

By the way, the process in which crust plunges back into the Earth is called subduction. Oceanic plate may subduct below continental plate. Old ocean floor may then become molten rock. When the two parts of ocean floor move away from the midocean ridge, lava hardens to form new ocean floor.

Once again we get persuaded that the slow and violent movements of the earth crust, though divided by the classification, are just the different phenomena of the same essence.

The areas of the Earth's crust can be moved up and down through faulting, folding and uplifting. But there is another process in which the crust moves up and down. Beneath the Earth's crust is a layer called the mantle, extending downward about 2900 kilometers to the Earth's core. The mantle is made of rock that flows slowly-sort of like malasses or hot, thick tar. Because the mantle is much denser than the crust, the solid, rocky crust floats on the mantle. The floating crust exerts a downward force on the mantle. But the mantle also exerts a force. A balance exists between the downward force of the crust and the upward force of the mantle. The balancing of these two forces is called isostasy. If material is added to an area of the crust, that area will float lower on the mantle. If material is removed, that area will float higher. The crust is always balanced on the mantle. Isostasy explains why some low-laying regions – such as Norway, Sweden, and Finland – have slowly risen. Thousands of year ago, these northern European countries were covered by tons of ice. The melting of the ice removed material from the crust. The land began to float higher on the mantle. Crustal rock can also sink. The Mississippi River has dropped millions of tons of mud and sand particles into the Gulf of Mexico. The addition of materials of the crust of the Gulf floor causes it to sink. But the depth of the water in the Gulf has not changed. A balance is maintained between the building up and the sinking of the Gulf floor.

Slow crust movements are known as secular movements, secular meaning lasting for centuries. Beside Scandinavian Peninsula, still rising at the rate of one centimeter a year, other coastal regions experience sinking. Evidence of this are the flooded ruins of ancient houses, the extensions of river valleys now under water, etc. Portions of the North Sea coast are gradually sinking at the rate of 10 cm per hundred years.

Movements of the earth crust occur everywhere, as has been proved by precision instruments. Some regions of the Donets Hills (Донецький кряж) and the Central Russian Upland (Середньоруська височина) in the East European Plain (Східноєвропейська рівнина), for example, are gradually rising. Moscow and its neighborhood, on the other hand, are gradually sinking at the rate of approximately several millimeters a year.

In the course of thousands of years, secular earth movements have greatly changed the shapes of the continents. When land rises and its area increases, islands are sometimes connected with the mainland. When land sinks and is flooded by the sea, the mainland is sometimes divided into parts and new islands are formed. Thus, several thousands of years ago the British Isles were part of the continent. Many regions, where there is land today, were formerly the bottoms of seas, e.g., the Central Danubian and Lower Danubian Plains.

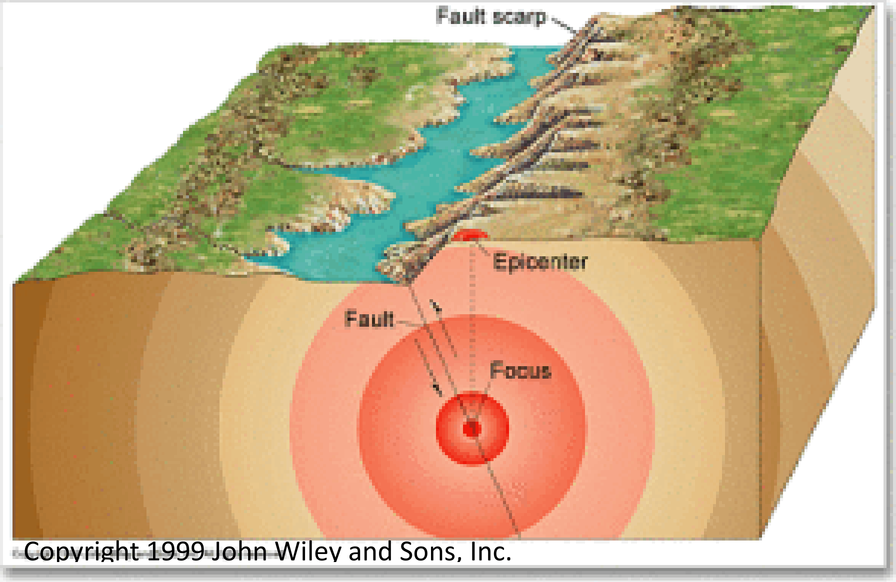
2.2.2. Earthquakes

An earthquake is the shaking and trembling that results from the sudden movement of part of the earth’s crust. The earthquakes are caused by the internal forces of the earth which act violently. When rocks in the earth’s crust break, earthquake waves travel through the earth in all directions. The ground shakes and trembles. During a severe earthquake, the ground can rise and fall like waves in an ocean. The motion of the ground causes buildings, trees, and telephone poles to sway and fall. Loud noises can sometimes be heard coming from the ground.

More than one million earthquakes occur every year. This is approximately one earthquake every thirty seconds. The vast majority of earthquakes are so small that the surface of the earth barely moves. But about twenty earthquakes a year cause severe changes in the Earth’s surface. They cause serious damage to buildings and dramatic loss of life in populated areas.

The most common cause of earthquakes is faulting. Rocks break and slide past one another. Energy is released during this process. As the rocks move, they cause nearby rocks to move also. The rocks continue to move in this way until the energy is used up.

As was already stated, earthquakes occur at converging or transform plate margins. As two plates move towards each other, one can be pushed down under the other one and into the mantle. If this plate gets stuck it causes a lot of strain in the surrounding rocks. Sideways-moving plates can also get stuck. When this tension in the rocks is finally released it produces strong shock waves known as seismic waves. The shock waves spread out from the focus – the point where the earthquake starts. Near the focus the waves are stronger and cause more damage. The epicenter is the point on the Earth's surface immediately above the focus.



Copyright 1999 John Wiley and Sons, Inc.

Fig. 9. The relationship between the epicenter and the focus of an earthquake.

The San Andreas Fault in California extends near the border with Mexico to the south through the city of San Francisco and continues on and off shore to the coast of northern California. It is 960 kms long and 32 kms deep. The land to the west of the San Andreas Fault is slowly moving north. The land to the east of the fault is slowly moving south. But the rocks along the fault do not all move at the same time. Earthquakes occur in one area and then in another. One of the worst of the disasters occurred in 1906, when movement along a small section of the San Andreas Fault caused the famous San Francisco earthquake. Here are the examples of major recent earthquakes:

Place	Year	Deaths	Size
Seattle	2001	0	7.2
India	2001	20,000	7.7
Turkey	1999	10,000	6.7
Kobe-Japan	1995	5,000	7.2

Earthquakes also occur on the floor of the ocean. They often produce giant sea waves called tsunamis. Tsunamis can travel at speeds of 700 to 800 kms per hour. As they approach the coast, tsunamis can reach height of greater than 20 meters. It can cause great damage.

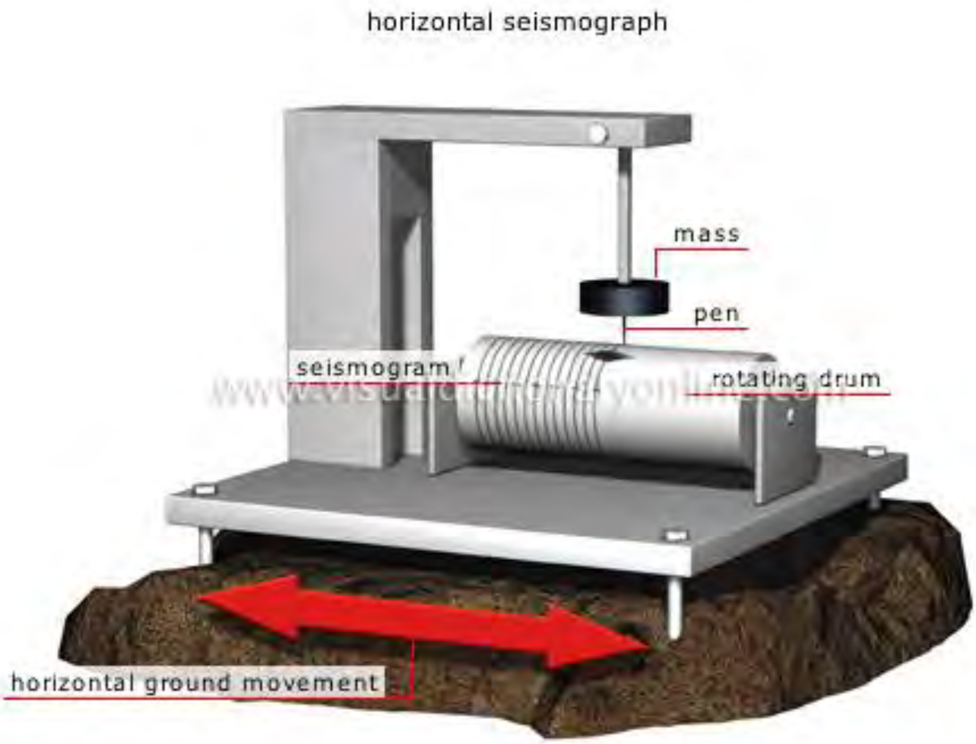
Earthquake waves are known as seismic [ʻsaizmik] waves. There are three main types of seismic waves. Each type of wave has a characteristic speed and manner of travel. Seismic waves that travel the fastest are called primary waves, or P waves. P waves arrive at a given point before any other type of seismic wave. P waves travel through solids, liquids, and gases. They move through the Earth at different speeds, depending on density of the material through which they are moving. In denser material, their speed is greater. P waves are push-pull waves. As P waves travel they push rock particles into the particles ahead of them, thus compressing the particles. The rock particles then bounce back. They hit the particles behind them that are being pushed forward. The particles move back and forth in the direction the waves are moving.

Seismic waves, that do not travel through the earth as fast as P waves do, are called secondary waves, or S waves. S waves arrive at a give point after P waves do. S waves travel through solids but do not through liquids and gases. Like P waves, S waves speed up when they pass through denser material.

Part of the Earth's interior is molten, or a hot liquid. Because S waves do not travel through liquids, they are not always recorded at all locations during an earthquake. S waves cause rock particles to move from side to side. The rock particles move at right angles to the direction of the waves.

The slowest-moving seismic waves are called surface waves or L waves. L waves arrive at a given point after primary and secondary waves do. L waves originate on the Earth's surface at the epicenter. Then they move along the Earth's surface the way waves travel in the ocean. Just as the water's surface rises and falls with each passing wave, the Earth's surface moves up and down with each L wave that passes. L waves cause most of the damage during an earthquake because they bend and twist the Earth's surface.

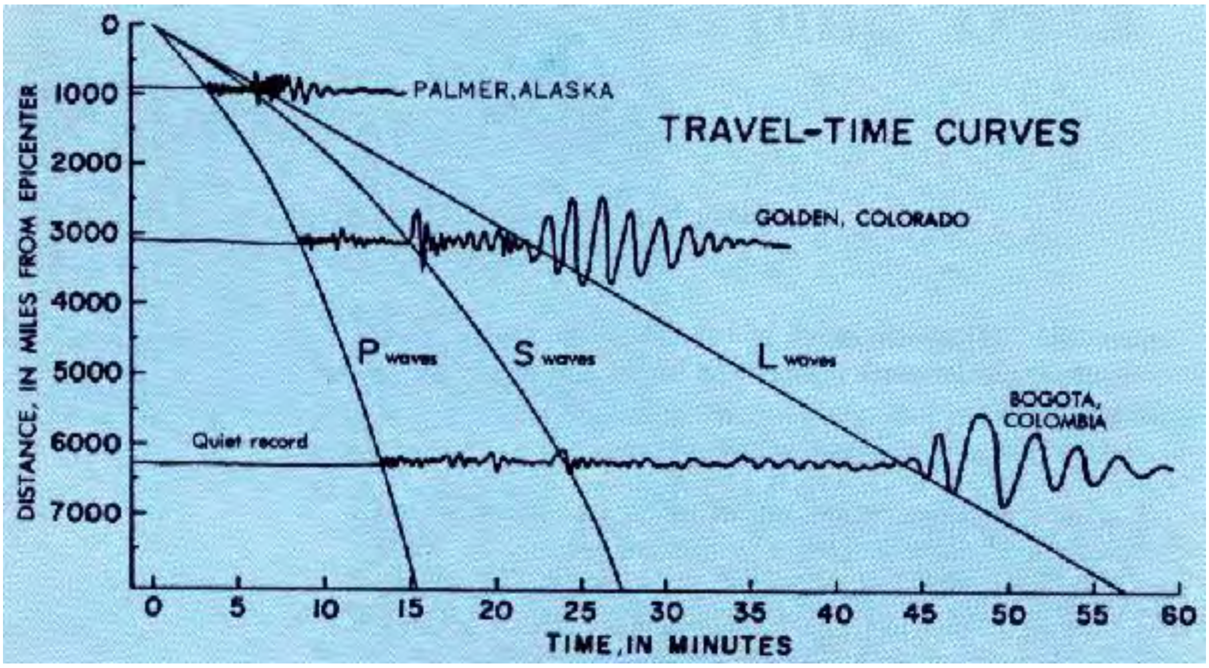
Shock (or seismic) waves are recorded by a sensitive instrument called a seismograph. Although crude seismographs were in use hundreds of years ago, the first practical seismograph was developed by John Milne in 1893. The figure below represents a seismograph:



Source: www.visual.merriam-webster.com/earth/geology/earthquake/seismographs_2.php

Fig. 10. Seismograph.

After-shocks, usually decreasing in strength but continuing over a period of several hours or days.



Source: www.earthquake.usgs.gov/learn/topics/seismology/keeping_track.php

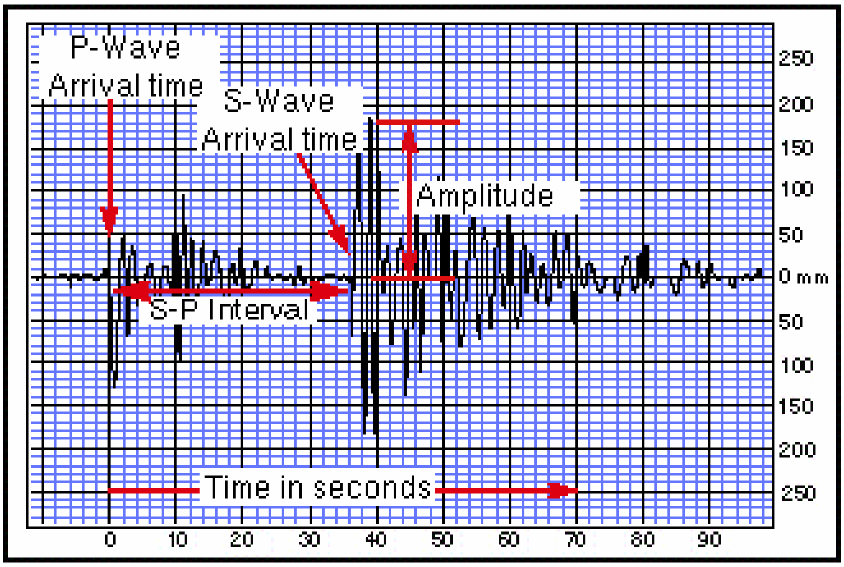
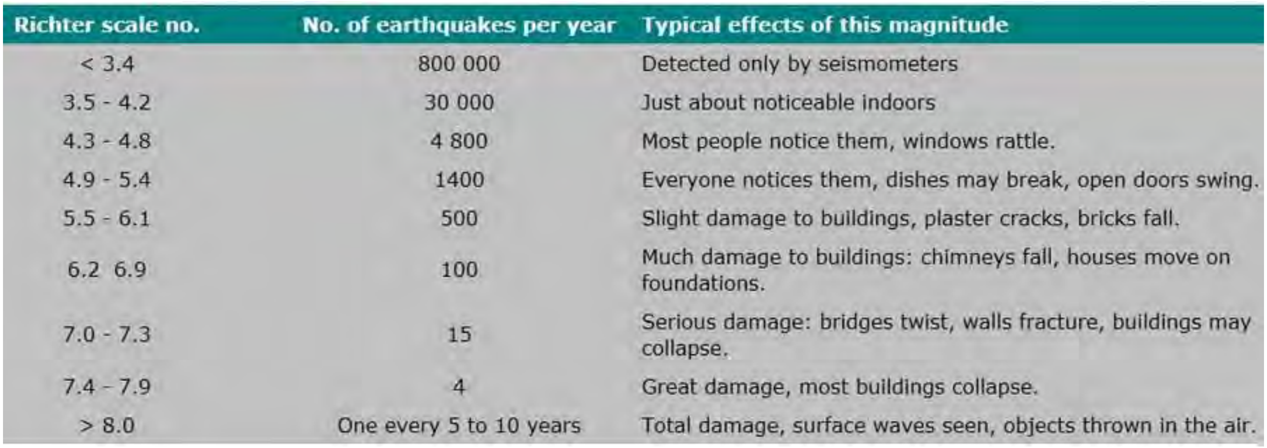


Fig. 11. A Seismograph Reading.

The strength of an earthquake is measured on the Richter scale. The Scale can, at first glance, be misleading because each point is ten times greater than the previous one. A seismometer is a machine with a seismograph on a revolving drum. Earthquake vibrations are recorded by a sensitive arm with a pen at the end which moves up and down. The Richter scale is for energy released, which is an open-ended scale. This is a logarithmic scale – which means that an earthquake with a score of 5 is ten times as powerful than one with a score of 4. Most serious earthquakes are in the range of 5 to 9. The earthquake in San Francisco in 1906 was the most powerful recent one, with a value of 8.6. Below is a popular form of the Richter Scale:



Richter scale no.	No. of earthquakes per year	Typical effects of this magnitude
< 3.4	800 000	Detected only by seismometers
3.5 - 4.2	30 000	Just about noticeable indoors
4.3 - 4.8	4 800	Most people notice them, windows rattle.
4.9 - 5.4	1400	Everyone notices them, dishes may break, open doors swing.
5.5 - 6.1	500	Slight damage to buildings, plaster cracks, bricks fall.
6.2 6.9	100	Much damage to buildings: chimneys fall, houses move on foundations.
7.0 - 7.3	15	Serious damage: bridges twist, walls fracture, buildings may collapse.
7.4 - 7.9	4	Great damage, most buildings collapse.
> 8.0	One every 5 to 10 years	Total damage, surface waves seen, objects thrown in the air.

A collision of plate boundaries, when two plates of continental crust moved towards each other, like Indo-Australian Plate moves towards the Eurasian Plate, caused serious earthquakes in India (1993), Iran (1997), and Afganistan (1998).

The Pacific Plate (oceanic crust) moves towards the Eurasian Plate (continental crust) causing earthquakes, e.g. Kobe (Japan) 1995.

The Pacific Plate moves more quickly than the North American Plate. Pressure builds up along the San Andreas Fault causing earthquakes in San Francisco in 1906 and 1989.

The amount of damage caused by an earthquake depends on several different factors. The earthquake's strength, the kind of rock and soil that underlies an area, the population of the area, and the time at which the earthquake occurs all influence how damaging a particular earthquake is. It is very important just how prepared countries are. Less economically developed countries (LEDCs) have less time, money and expertise to prepare for hazards. More economically developed countries (MEDCs) are better prepared but they still can't stop disasters happening. They can just limit the damage. In 1975, Chinese scientists predicted with great accuracy that an earthquake would occur in their country. Most of the people in three areas of the country were evacuated before the earthquake struck. Many thousands of lives were saved. And although it sounds a bit unscientific, some scientists in China believe that changes in the behavior of certain animals might help to predict earthquakes.

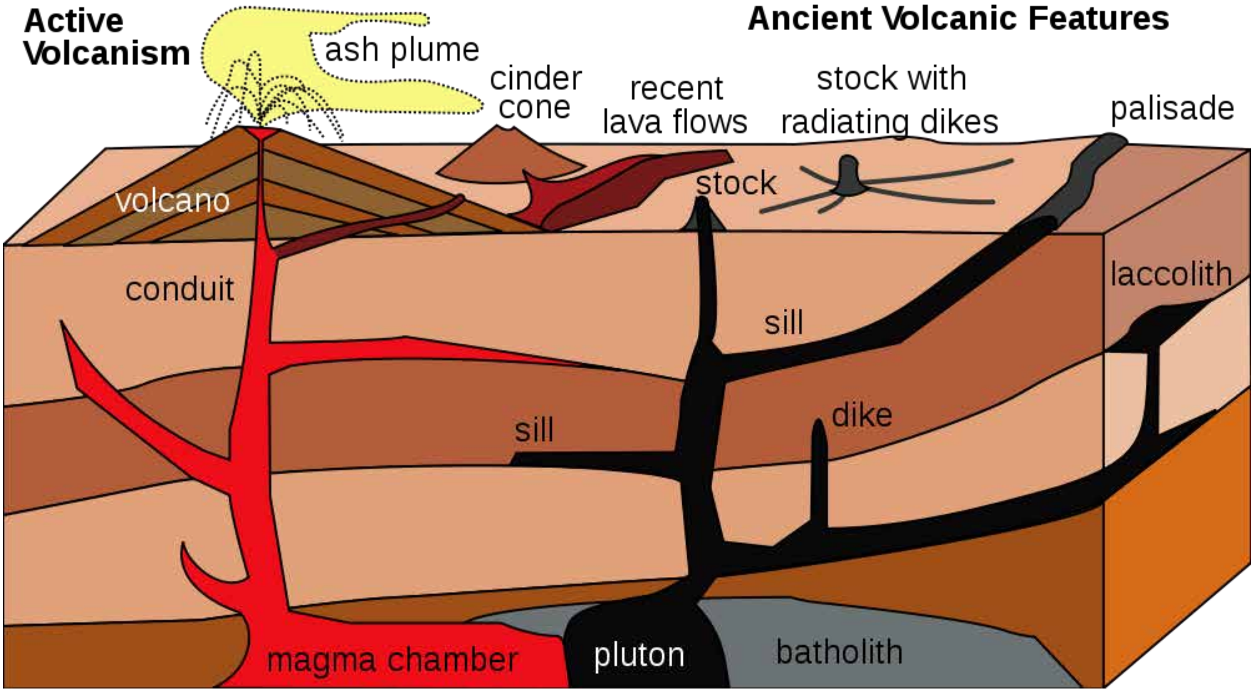
2.2.3. Magmatism

The substance, forming the earth's crust and the geospheres, according to some views, is in solid state, in spite of high temperatures at great depths. The pressure at these depths "maintains" the substance in solid state. Durable development of our planet favored the establishment of thermodynamic equilibrium in the earth crust and the geospheres below it. Any disturbance of this equilibrium (like the growth of temperature or lowering of pressure) melts the substance in some local places and it becomes liquid. This process is accompanied by the growth of volume. Magma (hot liquid rock) moves towards the weakened places of the earth's crust or flows onto its surface.

The penetration of magma into the crust is known as intrusive magmatism, or plufonism (Pluto being a god of the Earth's interior in ancient Greece). If magmatic materials are poured out upon the surface of the earth, as when lava flows from a volcano, this occurrence is known as extrusive or eruptive or effusive magmatism (Latin effusion meaning pouring out). It is also known under the name of volcanism (Volcano being a god of fire in ancient Rome).

Intrusive magmatism is the movement of magma rising up in the earth's crust but not being able to push itself to the very surface. Very often, such magma gradually hardens in the crust, forming various intrusions. Their shape is much dependent on surrounding rocks. The growth of temperature or the lowering of pressure in local spots of the mantle may cause the melting of its solid material. The density of silicate melt is lower than that of igneous or metamorphic rocks and it may reach the values more typical for the sedimentary rocks. The gases in magma make it more motile. This favors its penetration into the earth's crust as well as its eruption onto the surface.

Intrusive igneous rocks occur when magma cools very slowly. They have a course texture and form features like batholiths and tors. If magma flows along rock openings before cooling, it forms sills and dikes (Fig. 12).



Source: United States Geological Survey - Geologic History of Lake Mead National Recreation Area

Fig. 12. Forms of Intrusions.

When the magma cools very-very slowly, large hexagonal columns form – e.g. the Giants Causeway in Northern Ireland. The body of a batholith is composed mainly of granite or granodiorite. Their peripheral parts are composed mainly of syenite or diorite. Stocks are magmatic bodies which look like batholiths but are smaller, up to 200 kms^2 in area. The sills may have thickness from a few centimeters to several dozens or even hundreds of meters. Sheet sills dominate in layered surrounding rocks; they may be up to 45 meters thick and up to 150 kms long. They are composed mainly of basic and ultrabasic rocks. They spread out as essentially horizontal sheets between belts or layers of rocks.

Laccoliths are lens-like or mushroom-shaped intrusive bodies that have relatively flat undersurfaces and arched or domed upper surfaces. They are intruded between the bedding planes, are thicker in the center, and become thinner near their margins. A dome may be formed when rising magma causes the rock layers above it to fold upward. The uplifted area created by rising magma is called a dome. It is shaped roughly like the top half of a sphere. The outline of a dome is oval or circular. You can think of it as rather like a blister on the surface of the earth.

Over a long period of time, the uppermost rock layers may be worn away to reveal the dome's core of hardened magma. Domes that have been worn away in places form many separate peaks called dome mountains (The Black Hills of South Dakota).

Many of the domed mountains in the Crimea and Northern Caucasus (Beshtan, Mashuk, and Zalizna) have been formed by laccoliths. A dike is a tabular or wall-like mass of igneous rock that cuts across bedding planes when introduced into sedimentary rocks. Dikes commonly result from magma being injected into cracks and joints in the rocks, and range in size from a few feet to many miles in length. They are common in volcanic area.

During the solidification of magma, the amount of gases in it becomes excessive and they begin to evolve interacting with each other and with newly formed minerals. Some useful and rare minerals may be formed in this way. Generally speaking, the intrusive processes are very conducive to the formation of rare elements in hydrothermal and other veins. Everywhere on contacts with magma, different rocks form new minerals, not contained in either them or magma.

2.2.4. Volcanic Processes

When magma reaches the Earth's surface, it is called lava. The place in the Earth's surface through which magma and other materials reach the surface (and the magma becomes lava) is called a volcano. A lava flow is so hot that it incinerates every burnable thing in its path. In some places, lava can build up to form a cone-shaped mountain. Volcanoes are often (but not always) cone-shaped, formed by material from the mantle being forced through an opening in the Earth's crust, the vent. Volcanoes often have more than one vent. If there is more than one vent, lava will sometimes pour from the sides of a volcano as well as from the top.

All volcanic eruptions are not alike. Some are quiet, with lava slowly oozing from a vent. Other eruptions are very violent, with lava and other materials being hurled hundreds of meters into the air. Many scientists spend their working lives studying volcanoes, the "windows" into the interior of the Earth. They can determine the chemical composition of the magma from which the lava formed. Such data provide information about the composition of the part of the Earth that remains unseen. There are four main types of lava. One type is dark-colored and contains a lot of water. This lava is rich in the elements iron and magnesium. When it cools, igneous rocks such as basalt are formed. Another type of lave is light in color. It contains little water and is rich in the elements silicon and aluminum. Compounds of these elements account for its lighter color. On cooling, it forms the igneous rock rhyolite, which resembles granite. The third type of lava has a chemical composition which is intermediate between the two already mentioned types. Andesite is dominantly formed from this type of lava. The fourth type of lava contains large amounts of gases such as steam a carbon dioxide. When this lava hardens, it forms rocks with many holes in them. Pumice and scoria are igneous rocks formed from this type of lava. Some dark-colored lava is thin and runny and has a tendency to flow. The islands of Hawaii and Iceland were formed by many lava flows. But light-colored lava causes explosive eruptions. Because it is rich in silicon, it tends to harden in the vents of a volcano. Explosive eruptions are caused when lava in the vents hardens into rocks. Steam and new lava build up under rocks. When the pressure becomes great, a violent explosion occurs. During volcanic eruptions, many rock fragments are blown into the air. The smallest particles are called volcanic dust. Particles are less than 0.25 mm in diameter. Rock particles between 0.25 and 5 mm in diameter are called volcanic ash. Volcanic ash eventually forms some rocks. Both volcanic dust and volcanic ash can be carried away from a volcano by the wind. Larger rock particles are called volcanic bombs. They are a few centimeters to several meters in size. Small bombs about the size of golf balls are called cinders. Molten volcanic bombs harden as they travel through the air.

Volcanoes are extinct, dormant, or active: 1) extinct: will never erupt again; 2) dormant: has not erupted in 2000 years, and 3) active: has erupted recently and is like to erupt again. Mount Etna, Italy, has erupted in 2001.

Different types of volcanic eruptions form different types of volcanoes. Some volcanoes are built from quiet flows of thin, runny lava that spread over a broad area. Other volcanoes are formed from violent eruptions. Some volcanoes are formed from a combination of quiet flows of lava and violent eruptions.

Volcanoes made mostly of cinders and other rock particles that have been blown into the air are called cinder cones. They form from explosive eruptions. The material is loosely arranged so the cones are not high. But they have a narrow base and steep sides. Volcanoes composed of quiet lava flows are called shield volcanoes.

The lava flows over a large area because it is runny. Volcanoes built up of alternating layers of rock particles of lava (made up of lava and ash) are called composite volcanoes (Fig. 13).

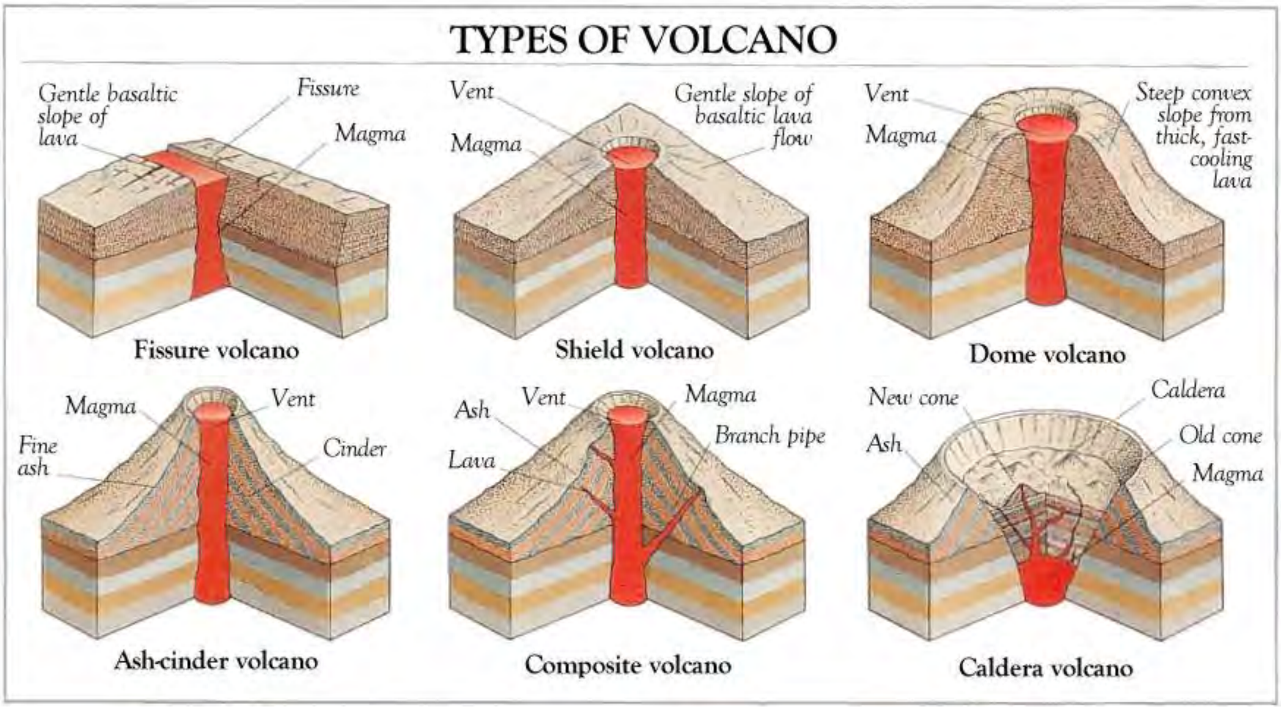


Fig. 13. Types of Volcanoes.

The most famous composite volcanoes are Mount Vesuvins near Neaples and Mount Etna in Sicily, both in Italy. Volcanoes are rather unpredictable phenomena. Some volcanoes erupt fairly regularly; others have not erupted within modern history. Most major earthquakes and volcanic eruptions occur in three zones of the world. There is a great deal of movement and activity in the Earth's crust in these zones. Volcanoes and earthquakes are near plate margins. There are over 600 active volcanoes in the world today. One major earthquake and volcano zone extends nearly all the way around the edge of the Pacific Ocean. This zone goes through New Zealand, the Philippines, Japan, Alaska, and along the western coasts of North and South America. The San Andreas Fault is part of this zone. This zone, that circles the Pacific Ocean, is called the Ring of Fire. A second major earthquake and volcano zone is located near the Mediterranean Sea. The Mediterranean Zone extends across Asia into India. Many countries in the zone, including Italy, Greece, and Turkey, have violent earthquakes. Many volcanic eruptions have also occurred in this zone. The third major zone extends through Iceland to the middle of the Atlantic Ocean. There is under the ocean a long range of volcanic mountains called the Mid-Atlantic Ridge. Volcano and earthquake activity in this area are due to the formation of new parts of the Earth’s crust along the ridge. Volcanic islands in the Atlantic Ocean, such as Iceland, are part of the Mid-Atlantic Ridge. The eruption of Vesuvins on August 24, AD 79 while it destroyed two lively cities of the ancient world, preserved evidence of these civilizations for almost two thousand years. Scientists still study a great historical treasure.

2.3. External (Surface) Dynamics Processes

The radiant energy of the sun finds its way into the atmosphere, hydrosphere, biosphere, the upper layers of the earth’s crust and becomes transformed into thermal, mechanical, chemical and biological energy. All these kinds of energy in interaction with the force of gravity cause the occurrence of different geological processes, which transform the Earth’s surface and the upper horizons of the Earth’s crust. These processes are known under the name of exogenic or external dynamics processes, whereas the changes in the land forms and the horizons of the upper portion of the Earth’s crust caused by exogenic processes are also known as exogenic or external geological phenomena and formations.

The most important agents of the external dynamics processes are:

- Solar radiation absorbed and accumulated on Earth and governing the thermal regime of its surface;
- Water getting on the surface and entering the surface layers and the Earth’s crust from the atmosphere;
- The gases of the atmosphere, particularly oxygen and carbon dioxide which evolve as a result of biochemical and mineral processes, and
- Plant and animal organisms.

Human beings are involved in all of these processes and perform their specific function on the Earth.

2.3.1. Weathering

It is the process whereby which the rocks undergo natural chemical and physical change at or near the surface of the earth (cortex of weathering). It is one of the most important of all geologic processes. It provides much of the material from which sedimentary rocks are formed and is important in the shaping of surface and responsible for the formation of soil. Rock fragments produced by weathering are removed by erosion – the loosening and carrying away of rock debris by natural agents. Weathering and erosion are constantly at work, wearing away the rocks of the earth’s surface.

Physical (mechanical) weathering takes place when a rock is reduced to smaller fragments without undergoing a change in chemical composition. This type of weathering, known also as disintegration, may be the result of a variety of physical forces. Frost action, alternate heating and cooling, freezing of water in cracks – all are the most potent agents of physical weathering. The most intensive physical weathering occurs under the influence of great gradients of diurnal and seasonal temperatures. In hot deserts the rock surface may reach the temperature of 60-70°C during the day and cool to nearly 0°C at night. Physical weathering becomes more intensive when water gets into cracks and fissures of rocks and creates capillary tension there. Freezing water increases in volume by one-tenth which produces a great pressure upon the walls of cracks.

In arid regions the same reaction is caused by soluble salts which crystallize from their solutions in cracks, fissures and other voids in the rocks. Anhydrite ($CaSO_4$), combining with water molecules, transforms into gypsum ($CaSO_4*2H_2O$) and increases its initial volume by one-third (33%). Here physics and chemistry go arm in arm. As rocks on the earth's surface undergo weathering, large pieces often break off the rocks. Over a long period of time, the rocks crumble and decay. One can see the results of weathering at the base of a mountain or on a mountain slope. Pieces of broken rocks pile up in these areas. These piles of rock fragments are called talus slopes. During mechanical weathering, rocks are broken into different shapes and smaller pieces. Sharp and angular rock fragments gradually become smooth and rounded. Often the pieces break off in curved sheets or slabs parallel to the rock's surface. This type of breaking off of rock is called exfoliation. Wind-blown sand causes mechanical weathering of rocks by abrasion. Abrasion is the wearing away of rocks by solid particles carried by wind, water, or other forces. In some areas of the Earth, particularly in certain mountain regions, the rocks are subjected to drastic temperature changes almost daily. Rocks of high mountain peaks expand, as they are heated in the day time, and constrict when subjected to freezing temperatures at night. In rocks with well formed distinct crystals of various minerals, possessing different coefficients of volumetric and linear expansion, the temperature variations cause the disintegration of a rock into the grains of the minerals composing it.

As a result of physical weathering, clastic sediments are formed. The solid rock becomes capable to retain and conduct through itself of water and air. Its surface area increases very considerably, which, in its turn, creates favorable conditions for chemical weathering. Chemical weathering, or decomposition, produces a chemical breakdown of the rock, which may destroy the original minerals and produce new ones. Chemical weathering produces rock materials that are basically different from the original rock. Although chemical changes occur in a variety of ways, the more common processes of decomposition are dissolution, hydration, hydrolysis, carbonation, oxidation, reduction, etc. Minerals can be added or removed from rocks. Most chemical weathering is caused by water and carbon dioxide. Water can form acids when it mixes with certain gases in the atmosphere. Carbon dioxide (CO_2), which is generally present in air, water, and soils, commonly unites chemically with certain rock minerals, greatly altering their composition. Substances, produced in this manner (carbonates and bicarbonates), are relatively soluble and therefore easily removed and carried away. Carbonic acid (H_2CO_3) is an effective agent in attacking such minerals as calcite and dolomite: $CaCO_3+H_2O+CO_2=Ca(HCO_3)_2$.

In nature carbonic acid is formed when CO_2 in the air dissolves in rain. This slightly acidic rain falls to the ground and is able to dissolve certain rocks on or beneath the surface of the earth. Sulfur oxides are a byproduct of the burning of coal. These compounds dissolve in rainwater to form sulfuric acid. Rain, that contains sulfuric acid, is one type of acid rain. It corrodes or wears away rocks and other materials rather quickly. The solubility of minerals in salt water also increases. Every increase in temperature by 10°C hastens the rate of chemical reactions 2-2.5 times. This explains a very intensive chemical weathering in equatorial areas of the globe and their slow passage in polar ones.

The reaction of feldspar with water illustrates the hydrolysis:

$$KAlSi_3O_8 + H_2O = KOH + HAlSi_3O_8.$$

An acid environment stimulates the reaction of water with minerals and is one of the most important weathering reactions:

$$\underset{\text{albite}}{4NaAlSi_3O_8} + 18H_2O + 4H^+ = \underset{\text{kaolinite}}{Al_4(OH)_8[Si_4O_{10}]} + \underset{\text{hydrous silica}}{8H_4SiO_4} + 4Na^+$$

In the reaction a primary mineral, albite, is converted to kaolin, a clay mineral. Some clay minerals tend to be resistant to further weathering. A process described above is sometimes called kaolinization. Chemical weathering of the most abundant primary minerals (orthoclase, albite, etc.) give birth to a number of secondary minerals like salts, oxides (hydroxides), and clay minerals. The kaolinite, even though resistant to further weathering, may be decomposed and disappear from the rock or soil:

$$Al_4(OH)_8[Si_4O_{10}] + 10H_2O = \underset{\text{gibbsite}}{4Al(OH)_3} + 4H_4SiO_4$$

Hydration of minerals (their chemical union with water) produces hydrous oxides:

$$\underset{\text{hematite}}{2Fe_2O_3} + 3H_2O = \underset{\text{limonite}}{2Fe_2O_3*3H_2O}$$

$$\underset{\text{anhydrite}}{CaSO_4} + 2H_2O = \underset{\text{gypsum}}{CaSO_4*2H_2O}$$

Oxidation occurs as oxygen, assisted by moist air, combines with minerals to form oxides. Rocks and minerals containing iron compounds are especially susceptible to this type of decomposition. Certain iron compounds (for example, pyrite) form acids when oxidized:

$$2FeS_2 + 2O_2 + H_2O = 2FeSO_4 + H_2SO_4;$$

$$12FeSO_4 + 6H_2O + 3O_2 = 4Fe_2(SO_4)_3 + Fe(OH)_3;$$

$$2Fe_2(SO_4)_3 + 9H_2O = 2Fe_2O_3*3H_2O + 6H_2SO_4.$$

The acid affects the rocks and thereby hastens the process of decomposition. The joint action of oxidation, hydration, and carbonation is a powerful agent of weathering of amphiboles and pyroxenes:

$$4CaFe[Si_2O_6] + O_2 + 4H_2CO_3 + 4H_2O = \underset{\text{calcite}}{4CaCO_3} + \underset{\text{goethite}}{4FeOOH} + \underset{\substack{\text{hydrous}\\ \text{silica}}}{8H_2SiO_3}$$

The weathering of certain primary silicate minerals is very important for the formation of parent materials and soils. Ferromagnesian silicate weathering may be represented by the generalized scheme:

$$\text{(Olivines Pyroxenes Amphiboles)} + O_2 + H^{+}Fe(OH)_3 + Mg^{2+} + Ca^{2+} + \text{silica residue.}$$

Soils dominated by these minerals in the fine silt and clay fraction are the minimally weathered soils all over the world, but mainly the soils of the desert regions where limited water keeps chemical weathering at its minimum.

Muscovite mica weathering proceeds in the following way:

$$\text{Muscovite}^{-K+} \rightarrow \text{Illite}^{-K+} \rightarrow \text{Vermiculite} \rightarrow \text{Smectite.}$$

Biotite mica, much more widespread in nature, yields in the following products:

$$\text{Biotite}^{-K+} \rightarrow \text{Chlorite}^{-Mg2+} \rightarrow \text{Vermiculite} \rightarrow \text{Smectite.}$$

Potassium cation released from micas and hydromicas (illite) becomes available for plant nutrition.

As to organic or biological weathering, some geologists are still undecided as to indentifying it among the types of weathering, as its mechanisms, in their essence, belong to either physical or chemical weathering. Anyway, biological weathering is very potent, and soil formation would be impossible without it. The activities of various organisms promote rock disintegration and chemical transformation of primary minerals into secondary ones. Tree roots, which frequently grow in rock crevices, can exert sufficient pressure to force rock fragments apart (root pry). Some bacteria (Bac. Extorquens) are capable to disintegrate and decompose (dissolve) feldspars. Lichens exude organic acids and decompose feldspars. The residues are no less efficient decomposers of aluminosilicates and iron and aluminum oxides. Some burrowing animals as a rodents, worms, and ants, bring to the surface rock particles to be exposed to the action of weathering. All these organic activities, taking place in the masses of disintegrated rocks, signify the beginning of the primary process of soil formation. The roots and residues of higher plants accelerate such process very much.

The upper part of the earth’s crust, where the processes of weathering are noticeably active, is sometimes called the cortex of weathering. The thickness of the cortex of weathering may be within the range of a few meters to one or two hundred of meters. It was developing for a different period of time. Some cortexes (corti) of weathering were developing since very ancient geologic epochs. They reflect the climatic conditions of weathering, according to which different types of cortex have been identified. Thus for taiga zone, it is reasonable to speak about siallitic acid type of cortex, as in these conditions, carbonates of bases have been leached from the cortex of weathering and it became ”unsaturated with bases”. Kaolinite and mica are two the most widespread minerals among the products of weathering.

Forest-Steppe and Steppe zones of Ukraine are characterized by siallitic saturated cortex of weathering, as the carbonates of bases were not leached from the cortex of weathering and parent materials of soils. Hydrous micas and montmorillonite clays are widely encountered in them. Calcium dominates among the exchangeable cations of soils. In humid climate of tropics, the cortex of weathering becomes enriched with the oxides and hydroxides of iron, aluminum and other elements. Montmorillonite clays and hydrous micas are not resistant to weathering under such conditions:

$$\underset{\text{pyrophyllite}}{Al_2(OH)_2[Si_4O_{10}]} + 10H_2O \rightarrow \underset{\text{gibbsite}}{2Al(OH)_3} + \underset{\text{hydrous silica}}{4Si(OH)_4}$$

Even kaolinite decomposes resulting in the formation of gibbsite and silicic acid:

$$Al_4(OH)_8[Si_4O_{10}] + 10H_2O \rightarrow 4Al(OH)_3 + 4Si(OH)_4$$

Gibbsite is a mineral more resistant to weathering then kaolinite. The loss of silicic acid by leaching results in the progressive loss of silicon. This type of weathering is called feralllitic. In arid climates physical weathering dominates as chemical one is impeded by the lack of water. B.B. Polynov identified a clastic type of the cortex of weathering in which the products of weathering remain in the place of their formation. Eluvium of a clastic rock formed from a solid rock is shown in the following figure (Fig. 14).

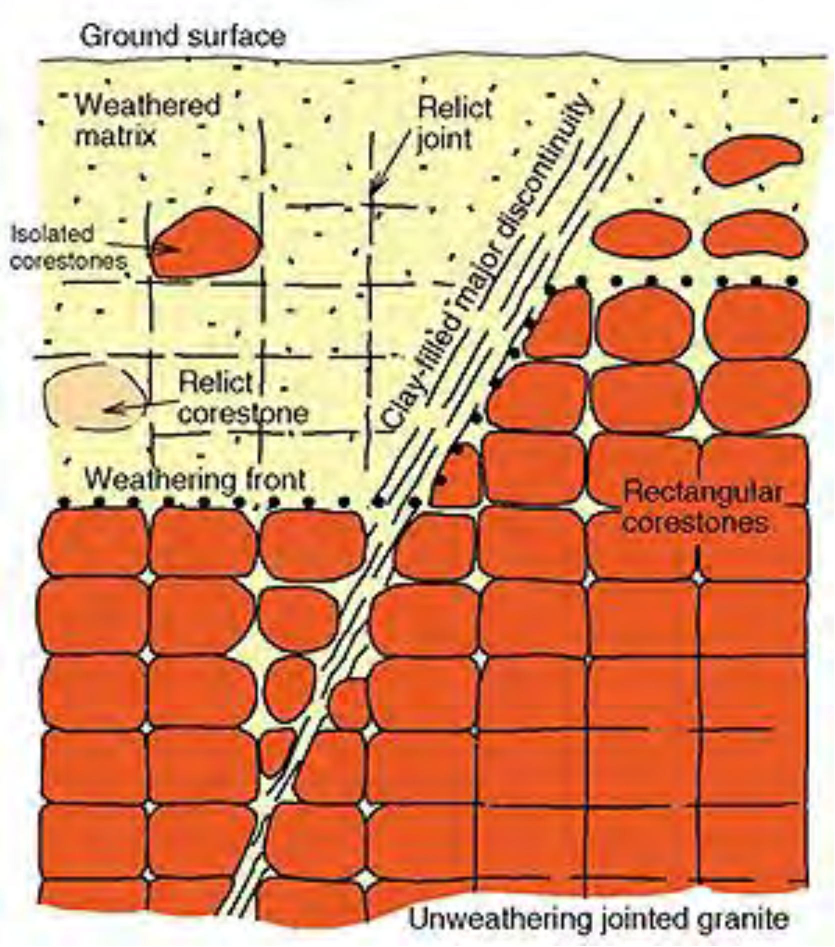


Fig. 14. Eluvial Cortex of Weathering.

Any process of weathering forms the products of two types:

- Removable or soluble products: salts, acids, bases, etc.
- Residual products not removed from the place of weathering and resistant to decomposition under given climatic conditions.

From the standpoint of soil science, the cortex of weathering is classified into two types: allitic and siallitic, siallitic being the dominant all over Ukraine, excepting probably some areas on the southern slopes of the main range of the Crimean Mountains (our subtropical zone). In tropics and subtropics the cortex of weathering is the thickest. Some scientists (including B. B. Polynov) identify the stages of the processes of weathering which depend upon climate and the composition of original rocks. According to B. B. Polynov, the stages are as follows:

1) clastic with the dominance of mechanical disintegration;

2) siallitic saturated with the leaching of chlorides and sulfates, and preservation of a bulk of $CaCO_3$; among the secondary minerals, hydromuscovite, sericite and moutmorillonite are dominant;

3) siallitic acid, characterized by deep leaching and acid reaction. Kaolinite and halloysite are the dominant clay minerals, and

4) final allitic (lateritic) cortex, occurring in hot climates with excess or sufficient humidity. Red-colored bauxites and iron oxides (limonite) dominate here among the secondary minerals.

It was V. V. Dokuchaev who first paid proper attention to the climatic (and general geographic) zonality of weathering and connected it with zonality of soils.

2.3.2. Geologic Activity of Wind

Air moves from an area of high pressure to an area of low pressure. Winds are formed by this movement of air from one place to another. There are two general types of winds: local winds and global winds. Local winds blow from any direction and usually cover short distances. Global winds blow from a specific direction and almost always cover longer distance than local winds. Both local winds and global winds are caused by differences in air pressure due to unequal heating of the atmosphere.

During the day, the air over a land area is often warmer than the air over nearby lake or sea. The air is warmer because the land heats up faster than the water. As the warm air over the land rises, the cooler air over the sea moves inland to take its place. This flow of air from the sea to the land is called a sea breeze.

During the night, the land cools off faster than the water. The air over the sea is now warmer than the air over the land. This warm air over the sea rises. The cooler air over the land moves to replace the rising warm air over the sea. A flow of air from the land to the sea, called a land breeze, is formed.

A major land and sea breeze is called a monsoon. A monsoon is a seasonal wind, as in Arabic it means ‘season’. During part of the year, a monsoon blows from the land to the ocean. During the rest of the year, it blows from the ocean to the land. When a monsoon blows from the ocean to the land, it brings in warm, moist air. This results in a rainy season with warm temperatures and huge amounts of rain. Monsoon winds are very common in Asia.

Global winds. Unequal heating of the Earth’s surface also forms large global wind systems. In areas near the equator, the sun is almost directly overhead for most of the year. The direct rays of the sun heat the Earth’s surface rapidly. The polar regions receive slanting rays from the sun, which do not heat the surface as rapidly as the direct rays do. So temperatures near the poles are lower than those near the equator. At the equator, the warm air rises and moves towards the poles. At the poles, the cooler air sinks and moves towards the equator. This movement produces a global pattern of air circulation. Global winds do not move directly from north to south or vice versa, because the Earth rotates, or spins on its axis, from west to east. The paths of the winds shift in relation to the Earth’s surface. All winds in the Northern Hemisphere curve to the right as they move. In the Southern Hemisphere, winds curve to the left. This shift in wind direction is called the Coriolis effect due to the rotation of the Earth. At the equator (0° latitude), surface winds are quite calm. These winds are called the doldrums. Any winds that do form are weak.

Meteorologists and weather observers use a wind vane to determine the direction of the wind on the Earth’s surface. A wind vane points into the wind. An anemometer is used to measure wind speed, which is expressed in meters per second.

Air pressure has a great effect on the weather. An area of low pressure that contains rising warm air is called a cyclone. In a cyclone, cooler air moves in to take place of the rising warm air. The air currents begin to spin. Winds spiral around and into the center of the cyclone. The winds move in a counterclockwise direction in the Northern Hemisphere. Cyclones usually cause rainy, stormy weather.

A high-pressure area that contains cold, dry air is called an anticyclone. Winds spiral around the out from the center of an anticyclone. In the Northern Hemisphere, the winds move in a clockwise direction. The weather caused by anticyclones is usually clear, dry, and fair.

A hurricane is a powerful cyclone (a low-pressure area containing rising warm air) that forms over tropical oceans. Hurricanes that form over the Western Pacific Ocean are called typhoons. The rapidly spinning rising air forms a doughnut-shaped wall of strong winds, clouds and rainfall. Inside the wall, the air is calm. This calm center is called the eye of the hurricane. Outside the eye, winds may reach speeds between 120 and 320 kilometers per hour (33-89 meters per second). A typical hurricane lasts about 9 days.

Tornadoes are also incredibly destructive. A tornado is a whirling, funnel-shaped cloud. It develops in low, heavy cumulonimbus clouds. The air pressure at the bottom of the funnel of swirling air is extremely low. When this low-pressure area touches the ground, it acts like a giant vacuum cleaner. Luckily, the diameter of an average tornado is only about 0.4 kilometer. The length of a tornado's path varies, but it averages 6 kilometers. A tornado in Nebraska tossed a 225-kg baby grand piano almost 400 meters across a corn field!

Geological work of winds takes place in all climatic zones but under the following conditions it is the most intensive:

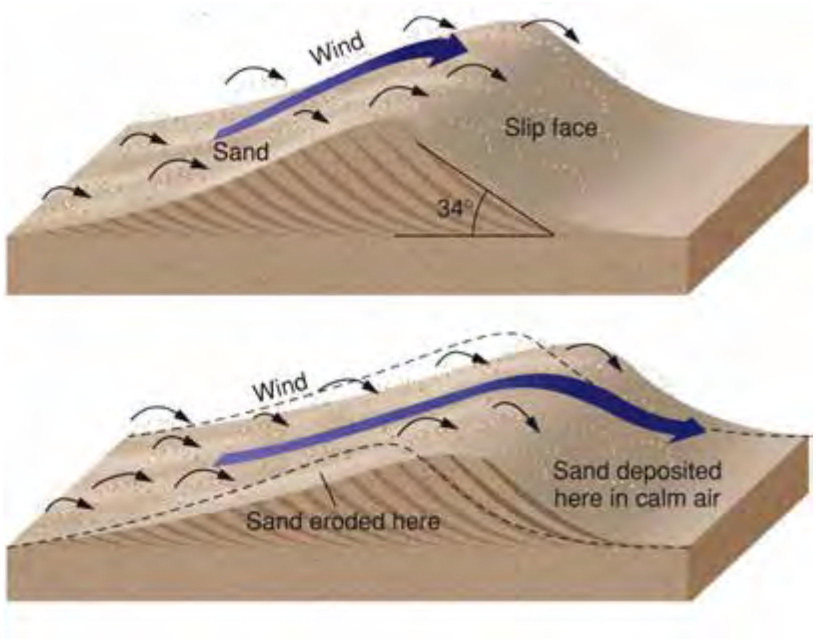
- arid climate with evaporation 6-10 times greater than precipitation;
- scarce vegetation cover with weak roots;
- intensive physical weathering producing much clastic material, and
- constancy of winds.

So the deserts and semideserts are the most susceptible areas for destruction by the wind. Such areas are about 20% of continental area.

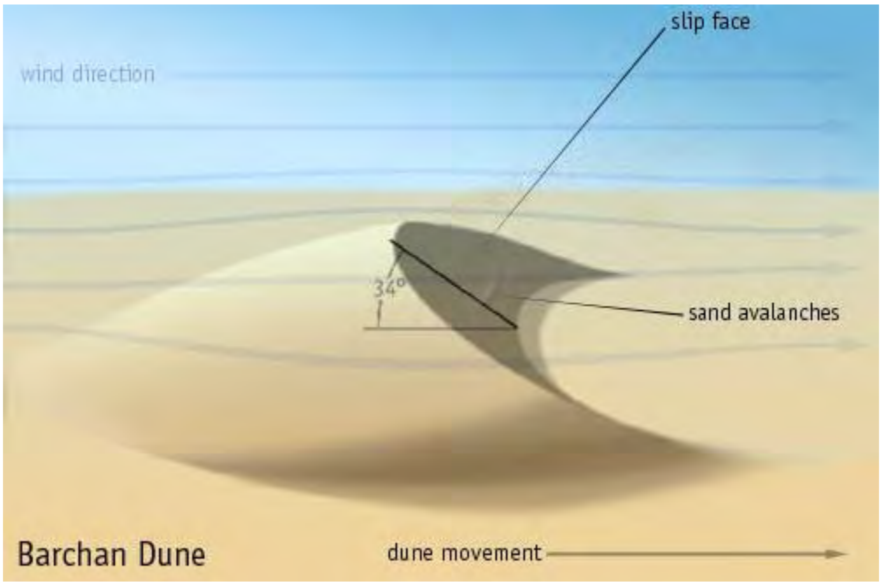
The destructive activity of the wind is when it wears land away. Wind is the most active agent of erosion in deserts; in plowed fields, and on beaches. In these places loose material is exposed at the earth's surface. This loose material can easily be picked up and carried by the wind. Wind erodes the earth's surface in two ways. Wind removes loose materials such as clay, silt, dust, and sand from the land. This type of wind erosion is called deflation. Wind erosion is usually considered a problem of dry land regions, but humid area sandy soils (our Polissya), muck soils (our overdried peats), and fine-textured soils, that are laid bare, can suffer severely from soil drifting. Soil carried by the wind by suspension (particles less than 0.05 mm in diameter), saltation (0.05-0.5 mm), and surface creep (larger than 0.5 mm). Wind erosion damage includes loss of soil, textural change, nutrient and productivity loss, abrasion, air pollution, and sedimentation. Annual soil losses higher than 700 mt/ha have been estimated for highly erodible, bare soils. The particles often act like a sandblaster, cutting and polishing rocks. This type of wind erosion is called abrasion. Wind has power to detach and transport soil grains. After transportation, particles and aggregates come to rest short or long distance from their origin. Detaching capacity of wind is directly proportional to the square of its velocity. Transporting capacity of wind is related to the third power of its velocity. The relationship between amount of soil removed from a unit area (the carrying capacity of the wind) appears to be proportional to the fifth power of its velocity. There is also an eolian transfer of soluble salts from salted areas to unsalted ones. The greatest concentration of solid material carried by the wind is in a layer of air not higher than 2 meters from the surface.

In many desert regions wind erosion forms wind caves by wearing away less resistant material. Sometimes wind erodes desert sands down to the depth where water is present. With water available on the surface, trees, shrubs, and grasses grow. Then a green, fertile area within a desert, called an oasis, forms.

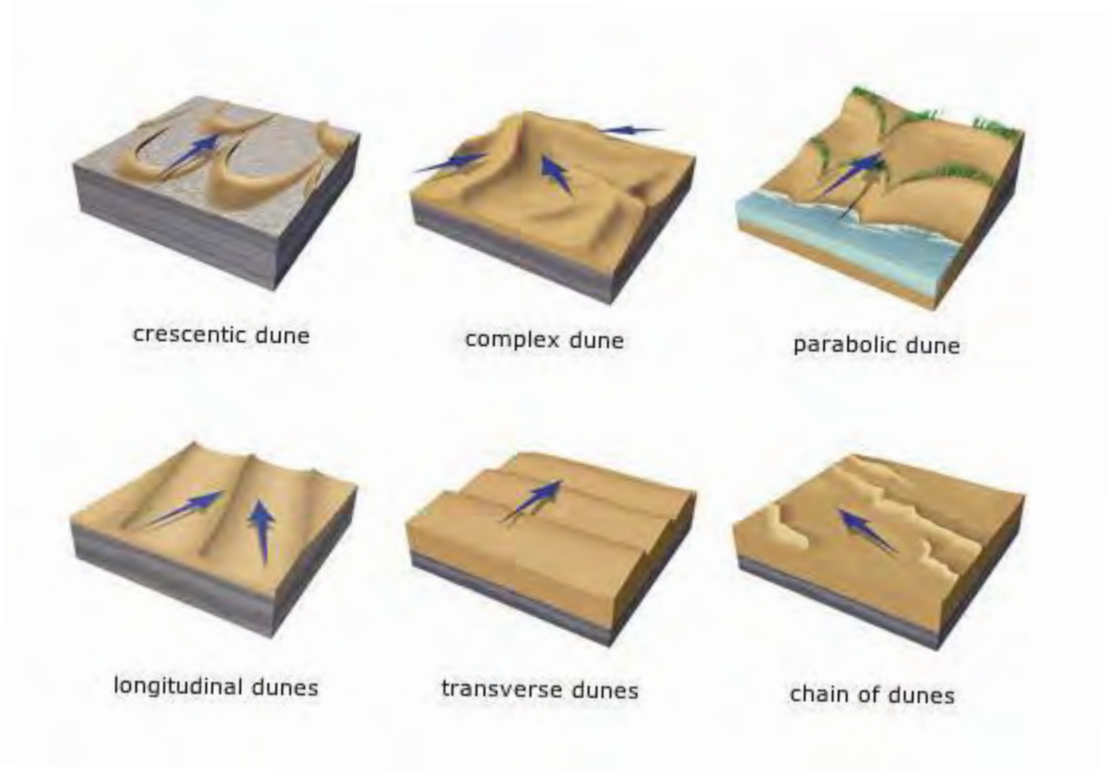
In desert areas and along shorelines, wind blown sand is often deposited near rocks and bushes. Wind flowing over these deposits is slowed down. More sand is deposited. The mounds of sand continue to grow to form sand dunes. A sand dune is a mound of sand deposited by wind. They vary in size and shape:



(by “Skull the Kepper”, from www.geocaching.com)



A. Movement of sand dune (from Wikipedia)



B. Aeolian dune shapes (by “Skull the Kepper”, from www.geocaching.com)

Fig. 15. A: Movement of sand dune; B: Aeolian dune shapes.

The side of the dune facing the wind has a gentle slope. Sand is carried up the gentle slope, or windward side, to the crest, or top of the dune. At the crest, the sand is dropped by the wind. The sand slides down the other side. This side of the dune, the slip face, has a steep slope. As the wind blows, sand dunes move across the area where they form. They move in the direction in which the wind is blowing. Sometimes they cover buildings, farmlands, and trees.

Some fine particles of silt are not deposited in dunes. They are deposited by the wind many kilometers from where they were picked up. Loess is formed. Loess deposits are very fertile. They are light in color and may be many meters thick. Three quarters of Ukrainian soils’ parent materials are represented by loess and loess-like loams. They are ideal parent materials. Among eolian forms of topography, in addition to dunes and barkhans, there also are loess plateaus and eolian knolls (small hills). Moving sands can be stopped by planting grasses with root systems well developed in a horizontal plane. The plantings of pine are also used for the purpose.

2.3.3. Geologic Activity of Precipitation

Atmospheric precipitations that reached the surface of the earth are distributed in the following way. Part of them, being evaporated comes back to the atmosphere; another part, having infiltrated through permeable rocks, is accumulated on the first from the surface water-tight layer forming ground water. The rest of precipitation is accumulated in temporary streams without beds which run along the lines of flow and join each other in larger flows with beds – channels and creeks which form what is called a river runoff. Streams cut valleys in all but the most arid landscapes. Most streams are integrated into river systems that carry runoff water to the ocean.

Sheet runoff. It is also called horizontal transfer of water. Surface runoff is when water flows over ground to rivers, lakes or the sea. Geological work of sheet runoff becomes more evident as the slope is steeper, rocks are more susceptible to detachment, and more rain or snowmelt water runs along the slope. Sheet erosion is the removal of thin layers of soil over the whole soil surface. Raindrop splash and surface flow cause sheet erosion, with splash providing most of the detaching energy and flow providing most of the transporting capacity. Sheet erosion is insidious because it is difficult to see. The first sign is when subsoil color begins to show, as cultivation mixes surface soil and subsoil. It is most apparent on upper portions of convex slopes. Gentle rains and light snows evaporate, sink into the ground, or are used by plants and animals. Heavy, fast downpours become runoff when the ground is saturated by long periods of rain, further precipitation becomes runoff. Steep slopes shed water quickly. Gentle slopes or flat areas hold water in place until it evaporates or sinks into the ground. Impermeable rocks cause rapid runoff.

Rill erosion. Rills are channels small enough to be obliterated by normal tillage operations. Rill erosion occurs when runoff water concentrates in streamlets as it passes downhill. This water has greater scouring action than sheet flow. It removes soil from the edges and beds of the streamlets. Rills frequently occur between crop rows and along tillage marks.

Gully erosion. Erosion channels too large to be erased by ordinary tillage are called gullies. The channels in deep loess soils are U-shaped with almost vertical walls. But V-shaped channels often develop where friable surface soils overlie cohesive, tight, nonerodible subsoils. Gullies are active when their walls are free of vegetation, and inactive when they are stabilized by vegetation.

Streambank erosion. Removal of soil material from the sides of running streams is called streambank erosion. It is usually greatest along the outsides of bends. Streams that are “unloaded” pick up sediment from their beds and banks.

Water pollution and sedimentation are the work of erosion too. Soil sediment is the greatest single pollutant of surface water. The muddying of streams and lakes reduces their value for home and industrial use. Rich bottomland soils owe their productivity to sediments eroded from the surface of upland soils. But they can be harmed if subsequent erosion deposits subsoil or other less productive material on them. The accumulation of sediment in the lower part of a slope is called deluvial process. Deluvium is a genetic type of continental deposits with slightly identifiable layering. In mountain regions, it is rich in clastic material. Speaking about erosion and sedimentation, it is necessary to remember that these processes may be natural or geologic and accelerated, or instigated by the activity of man. The normal rate of erosion under natural vegetation is in approximate equilibrium with the rate of soil formation. Accelerated erosion reduces the amount of plant growth a soil is able to support. A shallower soil, with its reduced capacity for storing water and plant nutrients, cannot match the productive potential of the uneroded soil.

Speaking about geological work of rainfall, it would not come amiss to remember a simple formula proposed by V. V. Slastikhin in 1964:

$$\Delta = i\sqrt{t}, \qquad (1)$$

where **Δ** is the force or work of a rainfall, **i** is its intensity in mm per minute, and **t** – its duration in minutes. What can be expected from a rainfall is given in the following table.

Table. 4. Rainfall Classification

Type	Strength (Δ)	Expected Effect
Drizzle	<1	No runoff; slight soil detachment owing to raindrop impact
Ordinary	1-3	Slight runoff; small soil loss
Moderate	3-5	Noticeable runoff; moderate soil loss
Average downpour	5-7	Water flows on the slopes; serious soil loss; small gullies possible
Severe downpour	7-9	Flooding of plainland valleys, severe soil loss, severe gullying
Very severe downpour	9-12	Floods on small rivers; catastrophic soil loss and gullying; active landslides

Rainfall water in alliance with the force of gravity on mountain slopes can cause catastrophic phenomena of mass movement, including talus slopes and piles, landslides, and mudflows. The assistance of rains may contribute to the growth of landslides, which are rapid movements of large amounts of material. Sometimes large blocks break away from steep mountain slopes. Landslides often follow long periods of rain. The gradual loosening of a mass of rock takes a long time. Landslides can carry millions of metric tons of rock to lower elevations causing tremendous damage. Mudflows also are rapid movements. These flows occur after heavy rains. When rains do come, the fine-grained particles wash down the slope. The mudflow follows old channels down to the valley below. There are thick mud spreads out into a cone-shaped mass. A mudflow can move whatever is in its path, including houses and farms. Some downslope movements are extremely slow. Creep is this kind of mass movement. Soil moisture helps to lessen the resistance of the fine particles to gravity.

2.3.4. Geologic Activity of Rivers

Rivers are the most important erosional agent because they affect so much area. Landscapes are largely the work of running water, even in desert regions. River systems begin when runoff follows the same channel rainfall after rainfall. Runoff follows the shortest path downward unless the material is too resistant for it to cut through. The flow then turns aside and cuts a path around the resistant obstacle. Once established, the same path is used by all later runoff. Small rills form first. They join to form creeks, which join to form streams, which join to form the main river. The network of channels is a drainage system. Many drainage systems have a treelike pattern. Rivers usually begin in mountains or hills. The downward pull of gravity gives them energy to cut away the land and form valleys. The larger streams are called tributaries. They flow into the main river. The main river is like the tree trunk. In time the main river empties into another river, a lake, or an ocean at a place called the mouth of the river.

The area, drained by a main river and its channels, is called a drainage basin. The land, that separates one drainage basin from another, is called a divide. But as the drainage system of a river develops, the divide becomes narrower. Sometimes a drainage system will cut through its divide and steal runoff from another drainage basin.

A drainage system grows larger by deepening its channels, widening its valleys, and adding more rills and gullies to its system. The river grows larger and faster, and the river valley grows deeper and wider. In time the river reaches a balance between the processes of erosion and deposition. One square kilometer of Ukrainian territory contains 0.25 kms of rivers. Of rivers over 10 kms in length there are 4000, of those over 100 kms – about 120, of those over 500 kms – eight. The largest rivers in Ukraine are named in the following table (Table 5):

Table. 5. The Largest Rivers of Ukraine

Name of a River	Enters into	Length (km)		Area of Basin, thousand km^2
		Total	Within Ukraine	
Dnieper	Black Sea	2201	981	504.0
Southern Bug	Black Sea		806	63.7
Psyol	Dnieper		717	22.8
Dniester	Black Sea	1362	705	72.1
Siversky Donets	Don	1053	672	98.9
Goryn’	Prypyat		659	22.7
Desna	Dnieper	1130	591	88.9
Inguletz	Dnieper		549	14.9
Vorskla	Dnieper		464	14.7
Sluch	Irpin’		451	13.8
Styr	Prypyat	494	445	13.1
Western Bug	Vistula	831	401	73.5
Teteriv	Dnieper		385	15.3
Sula	Dnieper		365	19.6
Ingul	Southern Bug		354	98.9
Ros’	Dnieper		346	12.6
Samara	Dnieper		320	22.6
Prut	Danube	967	272	27.5
Tissa	Danube	966	201	153.0
Seim	Desna	748	250	27.5
Zbruch	Dniester		244	3.4
Oskil	Siversky Donets	472	177	14.8
Danube	Black Sea	2900	174	817.0

Erosion basis for a given river is the level of the main river into which a given river falls. General erosion basis for any river is sea level. The Dnieper’s erosion basis is the level of water in the Black Sea. Life cycles of rivers are characterized by certain stages of their development in time. An immature river, or young river, is a river in an early stage of development. An immature river cuts a valley with steep sides into the earth’s surface. The valley is typically V-shaped, and river covers almost the entire valley floor. The waters of an immature river flow very quickly over rocks, producing rapids (пороги). Waterfalls are also commonly found in immature rivers.

A river that has been developing for many thousands of years is called a mature river. Because of continuous erosion, the rapids and waterfalls have largely disappeared. The river has also eroded much of the valley floor. The valley walls are far from the river itself. The floor of the valley is broad and flat. The course of the river has also become curved and winding, forming loops called meanders. A river reaches a balance between the processes of erosion and deposition. Then a river has a profile of equilibrium along its length. Drainage systems may be changed by any of the following:

- uplift or lowering of the headwater area where the river begins;
- uplift or lowering of the sea level;
- change in climate (the famous Russian geographer A. I. Voyeikov: “Rivers are always the product of the climate”), and
- wearing down of the divide (вододіл).

Speaking about the geological work of rivers, it is necessary to remember that a drainage basin works as a system (Table. 6):

Table. 6. Drainage Basin System

INPUTS	FLOWS	STORAGE	OUTPUTS
Precipitation	Surface run-off	Channel storage	River run-off
	Channel flow	Ground water storage	Evaporation
	Infiltration	Short-term storage	Transpiration
	Through flow	Soil moisture storage	
	Percolation	Vegetation storage	
	Ground water flow		

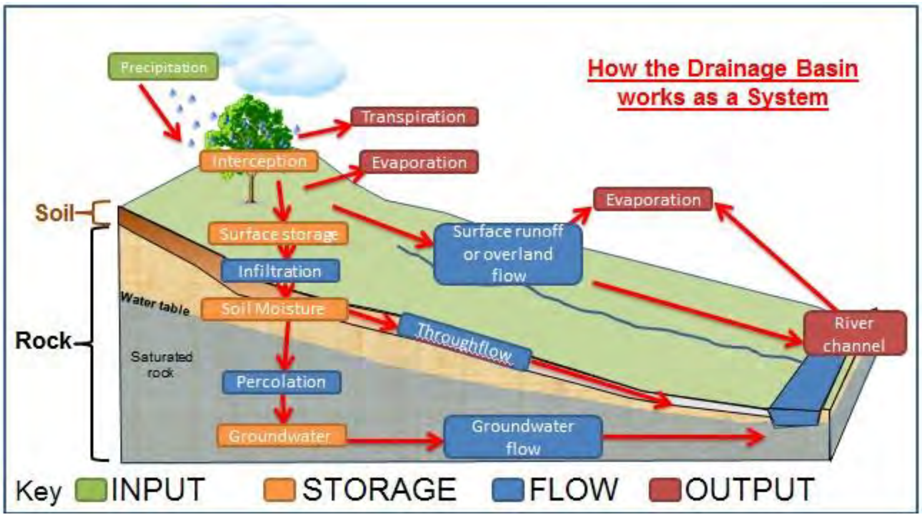


Fig. 16. Drainage Basin as a System (by Wendy Yin, Himani Deshpande, Momtafin Khan).

Water enters the drainage basin as precipitation. It goes through a series of flows and stores before reaching the seas as river run-off. The time between rain and river runoff varies with the basin characteristics (shape, size, rock type, vegetation). Energy is put into the system by the steepness of the hills/valley and the force of gravity. Water moves rock and soil material through the drainage basin system. It’s picked up when the water energy is high and deposited when the energy is low.

A river basin has several important features:

1) The source is where a river starts, usually in an upland area.

2) A tributary is a stream that joins the main river.

3) A confluence is the point where two rivers join.

4) The mouth is where the river flows into the sea.

5) An estuary is where the mouth is low enough to let the sea enter at high tide – this causes deposition, forming mud and sand banks which the river flows between.

A stream or river carries a large amount of sediments. In places where the stream or river slows down, sediments are deposited.

Oxbow lakes (старичні озера). Sometimes the meanders of a river form large, U-shaped bends. Erosion and deposition along such bends can cut these bands off from the river. A small lake called an oxbow lake is formed (oxbow = ярмо для вола):

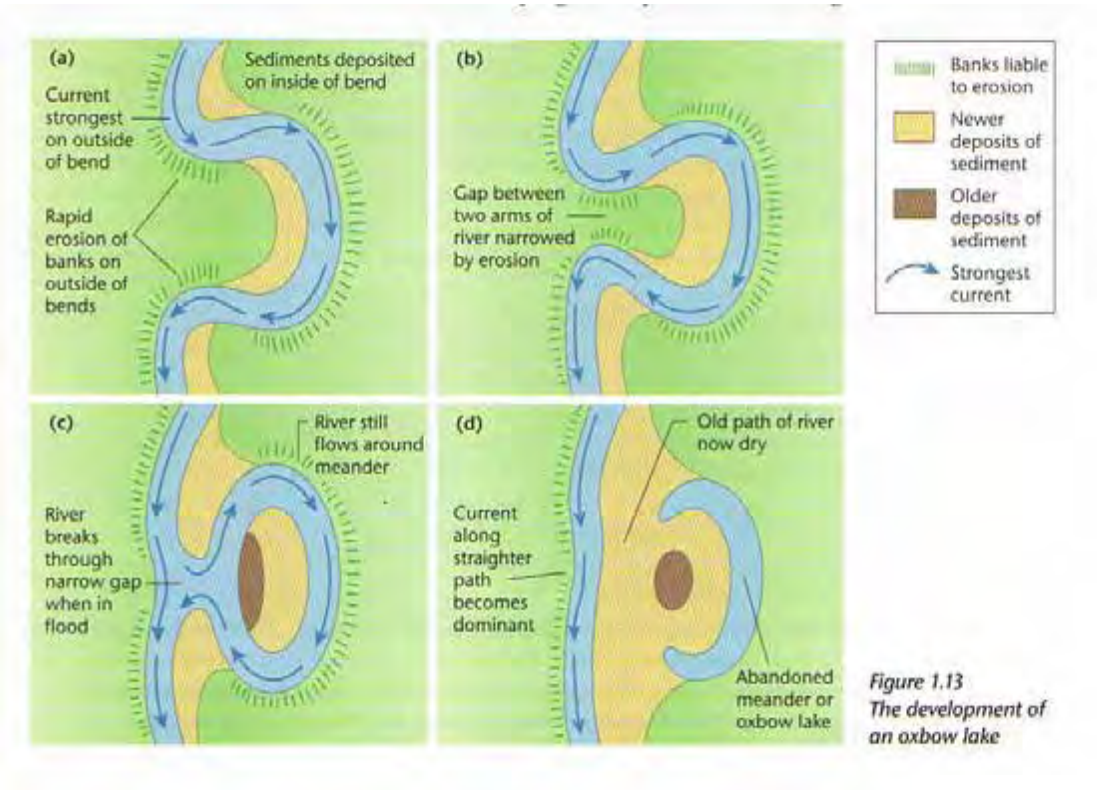


Fig. 17. An oxbow lake formation (by Weebly, from www.acege

Alluvial fans. When a river leaves the mountains and runs out onto a plain, its speed decreases. Nearly all the sediments the river is carrying are dropped. They build up to form an alluvial fan. The sediments spread out from the river channel in a fan like shape.

Deltas. Large amounts of sediments, deposited at the mouth of a large river that flows into a lake or an ocean, form a delta. A delta forms because the river's speed decreases as it runs into the body of standing water. The river cannot carry as much material when it is moving slowly. So it deposits much of the sediments. Sediments build up above the river's water level.

FLOOD PLAINS & LEVEES. On both sides of a mature river or stream, flat areas called flood plains form. After heavy rains or spring thaws, the river overflows its banks and covers the floor plain. Sediments are deposited on the plain. Repeated flooding causes sediment to build up. Flood plains have fertile soils. Sediments deposited on a flood plain usually consist of fine particles. The larger particles, which settle first, are deposited along the sides of the river. They form a ridgelike deposit called a levee (прирусловий вал). Rivers flow in linear features called valleys. Tectonic types of the valleys are given in the following figure:

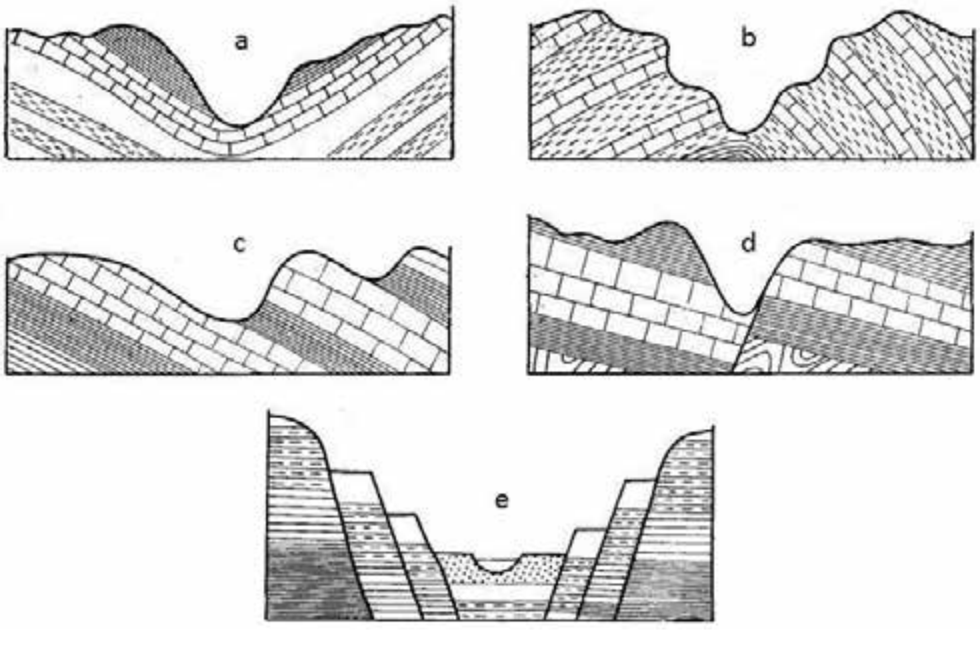


Fig. 18. Tectonic Types of River Valleys: a – synclinal; b – anticlinal; c – monoclinal; d – along the line of fracture; e – graven-valley.

The elements of a river valley are: riverbed, floodplain, slopes, terraces and banks. Terraces are formed when the base level of a stream changes and its channel is cut too deep for normal flooding to occur. The stream may have cut through a barrier to a lower base level, or uplift may have increased the slope of the stream channel. Either cause leaves the former flood plain higher than the stream and its new floodplain. The older floodplain becomes a terrace, but its origin is recognizable by its flat surface bordered by a slope down to the new floodplain on one side and a slope up to another terrace or the upland above on the other side. Terraces are subject to erosion as the new floodplain enlarges and to dissection where tributary streams cross them. Types of river terraces are given in the following figure:

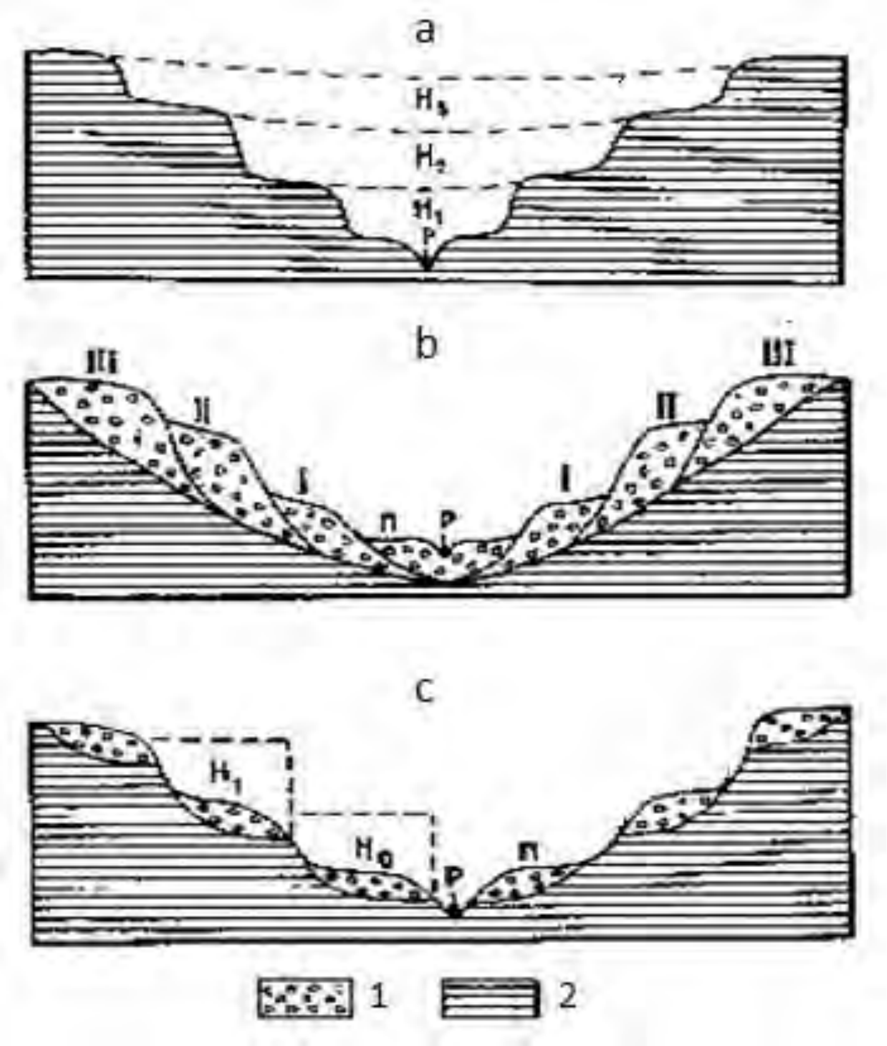


Fig. 19. Types of River Terraces:

a – erosional; b – accumulative; c – plinth-like (цокольні); P – riverbed; П – floodplain, I, II, III – above the floodplain terraces; H1, H2, H3 – erosion cycles, 1 – alluvial deposits; 2 – bedrock

A valley cross profile has three stages:

1) upper: near to the river's source the valley has a narrow floor and steep sides, i.e. is V-shaped.

2) middle: lower down the river, the floor is wider and the sides are more gently sloping.

3) lower: when the river is near to the sea it has a wide floor and gentle sides:

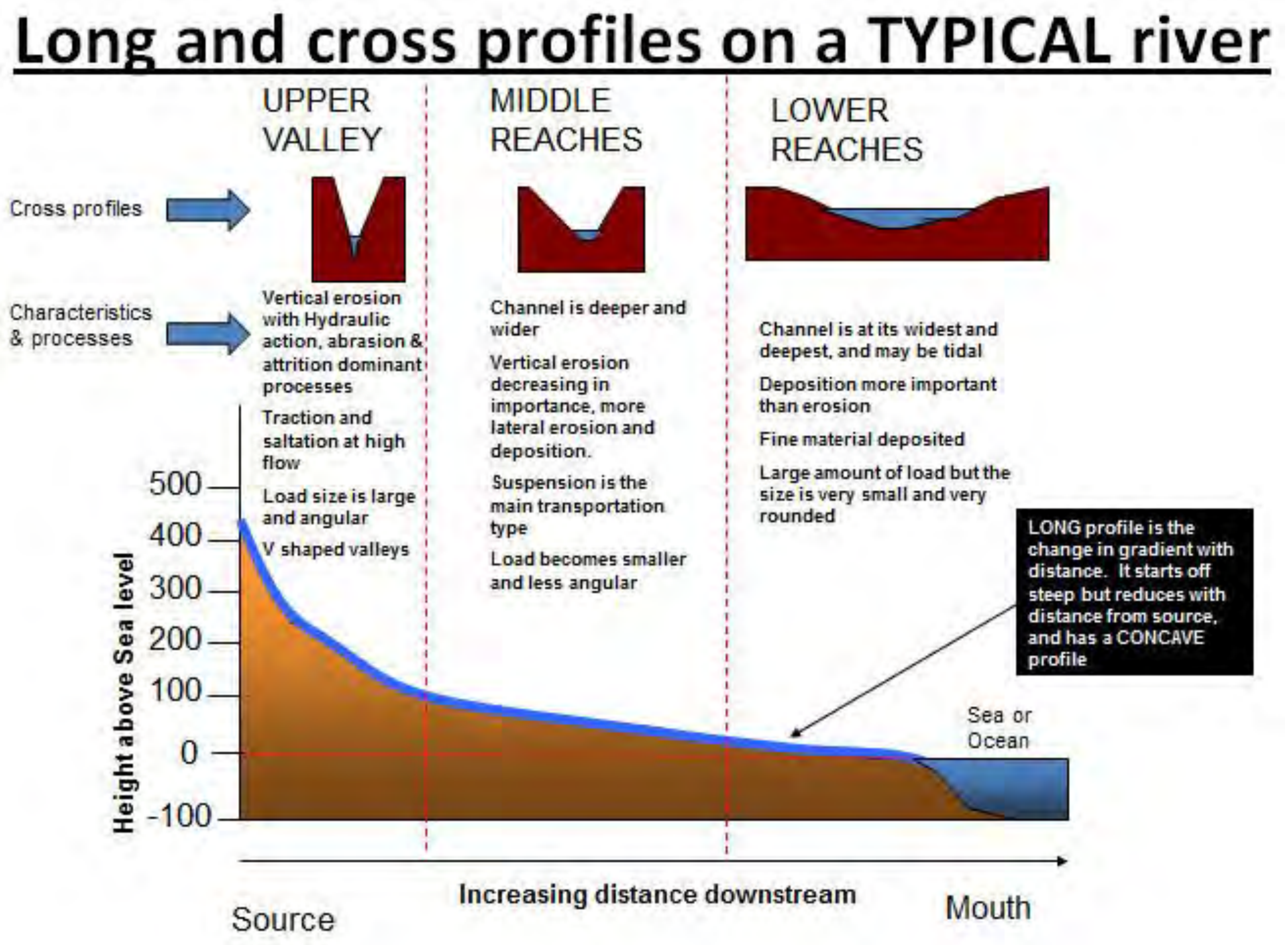


Fig. 20. Three Stages of Valley Development from Upper to Lower Reaches.

A river's long profile varies as it moves downstream:

1) In the upper stage, the river's gradient is quite steep.

2) In the middle stage it is more gentle.

3) In the lower stage it is very gentle and almost flat.

Rivers erode in four main ways, called erosion processes:

1) Corrasion or abrasion is when large pieces of bedload material wear away the riverbed and banks – e.g. in floods. If material collects in a dip, it swirls round forming a pothole.

2) Attrition means the rocks being transported are eroded. The sediment particles knock against the bed or each other and break apart, becoming smaller and more rounded.

3) Hydraulic action is when the force of the water wears away at softer rocks such as clay.

4) Solution or corrosion is when chalk and limestone dissolve in water.

River erosion is headward, vertical or lateral;

1) Headward erosion is when the furthest point upstream, the valley head, is worn away by rainwash, undercutting or soil creep. Soil creep is the slow movement of soil downhill over time.

2) Vertical Erosion deepens the valley as the water force grows – common in the upper stage when the gradient is steep.

3) Lateral Erosion widens the valley, combined with weathering of the sides – it's common in the middle and lower stage valley.

The movement of eroded material – transportation of the load occurs in four ways:

1) Suspension is when fine silt and clay material is carried along the water itself.

2) Saltation is when small sand-sized particles are bounced along the riverbed.

3) Traction is when larger materials like pebbles or boulders are dragged along the bed.

4) Solution is when eroded material dissolved in the water is carried away.

Deposition is when a river dumps its load. Material is deposited where the flow of the river is slower, forming what Ukrainians call прируслові відмілини. Interlocking spurs are caused by erosion. They are ridges produced when a river in the upper stage twists and turns round obstacles of hard rock along its downward pathway. These ridges interlock with one another like the teeth of a zip fastener.

Waterfalls are found at steep parts of the river bed. A layer of hard rock won't erode very easily so when the river reaches it, any softer rocks on the downstream side are eroded more quickly. The river bed gets steeper where it crosses the hard rocks and a waterfall forms. Waterfalls can form when the hard rock is horizontal, vertical or dips upstream. At the foot of the waterfall the water wears away the softer rock to form a plunge pool (водобійний колодець). Rapids (пороги) are a series of little waterfalls. They are found where there are alternative bands of hard and soft rock.

Estuaries are funnel-shaped river months. Most are found where an existing river has had its lower reaches flooded after changes in sea level. Deltas form when a river deposits silt too fast for the sea to remove it – because the sea is tideless or the load is too big. There are three main types of deltas (Fig. 21).

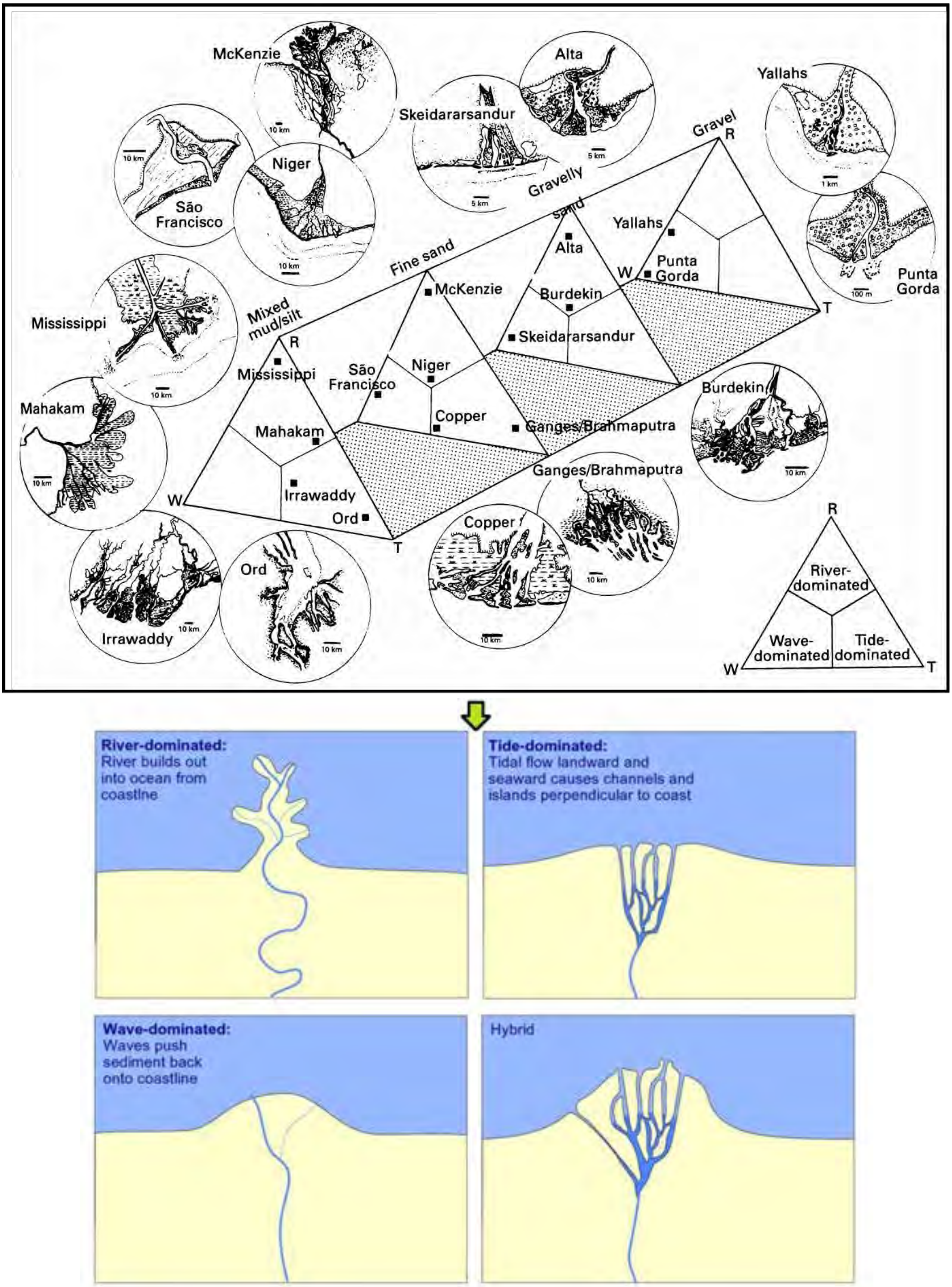


Fig. 21. Main Types of Deltas.

Now it is time to speak about the combined geological work of rainfalls and the rivers. In such a combination they may cause floods. Using the storm hydrograph, such major hazards can be predicted. The graph shows the change in the river discharge (volume of water flowing per second) over a short period of time after a storm. It is used to work out when a flood might be coming.

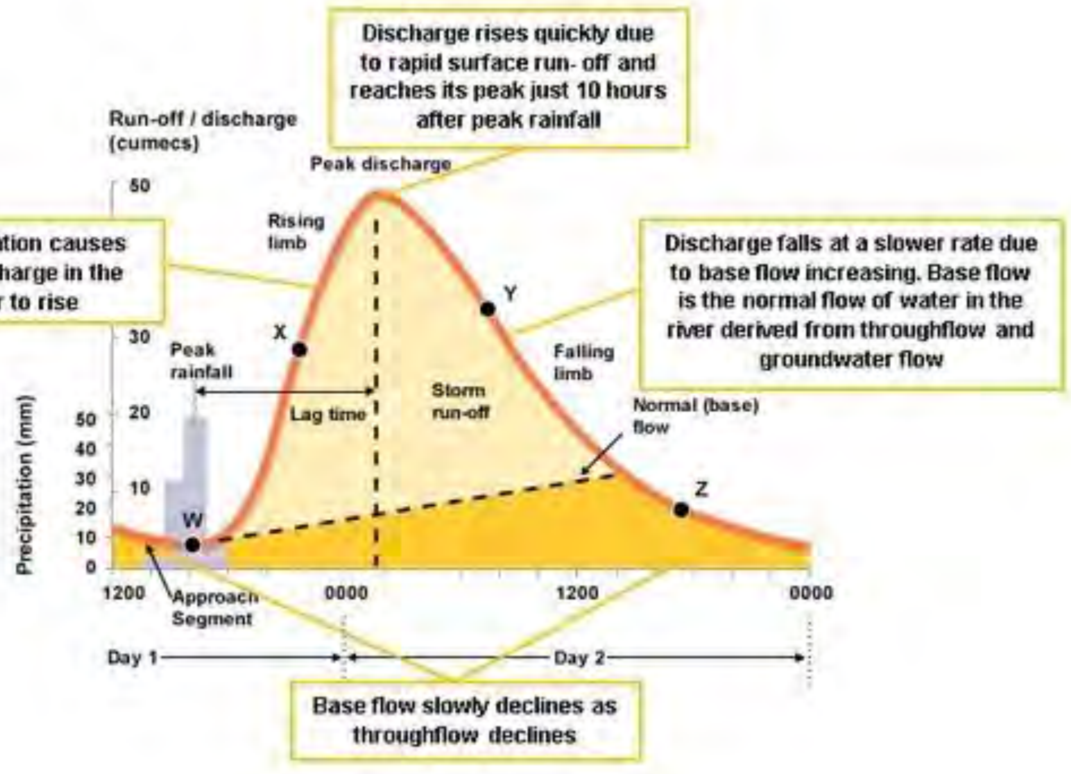


Fig. 22. Change in the river discharge over a short period of time after a storm.

1) The base flow is the normal discharge of the river.
2) The rising limb represents the increase in discharge after the storm.
3) The falling or recession limb represents the decrease in discharge.
4) The lag time is the amount of time between peak rainfall and peak discharge.

The river is likely to flood when the graph is steep. This is because there is a rapid increase in discharge over a short period of time and the river system is unable to transport it away.

Floods can cause extensive damage. Previous flood data can be analyzed to predict when a new flooding is going to hit. Changing land use can help reduce flooding. Afforestation of bare slopes in the upper reaches reduces run-off. Lag time is longer with less run-off or river discharge. Leaving land up river as pasture gives a continuous plant cover, reducing runoff. Scientists believe that the severe flooding we’ve had in recent years could be an effect of global warming.

2.3.5. Geologic Activity of Seas and Oceans

Two thirds of Earth's surface is covered by the ocean. The ocean is a continuous body of salt water that covers a little over 70 percent of the Earth's surface. Large as the ocean is, it is only part of the hydrosphere. Ocean water has a number of different elements in solution. Some substances are carried to the ocean by rivers that flow over weathered rock. As it was already stated, the ocean is a store house of dissolved minerals. Salt, magnesium, and bromine are three products extracted from seawater.

The power of the sea is also tremendous. Waves are energy movements through water caused by the wind. They are the main way in which the sea erodes, transports, and deposits material. The energy of a wave is determined by its height, which is the distance between its trough and crest, and length, the distance between two crests:

The geologic works of the wind and the sea cooperate to produce waves:

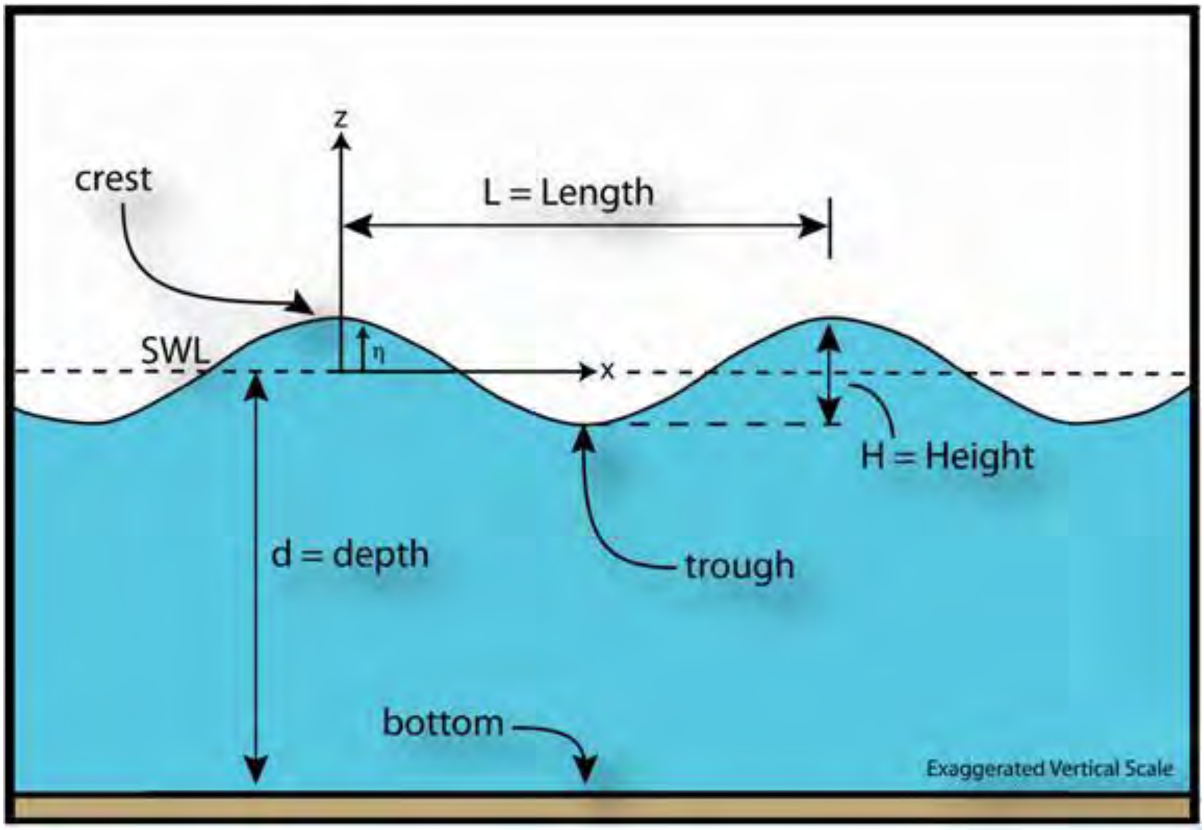


Fig. 23. Sea Wave Parameters (from "Highways in the Coastal Environment: Second Edition" by FHWA).

Wave height and length vary according to the speed and length of time a wave has been moving, and the fetch – the distance of open sea over which the wind has blown. Near the coast, waves slowdown in the shallow water, causing them to "break", becoming unstable. Seawater moving up to beach is swash, and moving back to the sea is "backwash". Waves can be constructive and destructive:

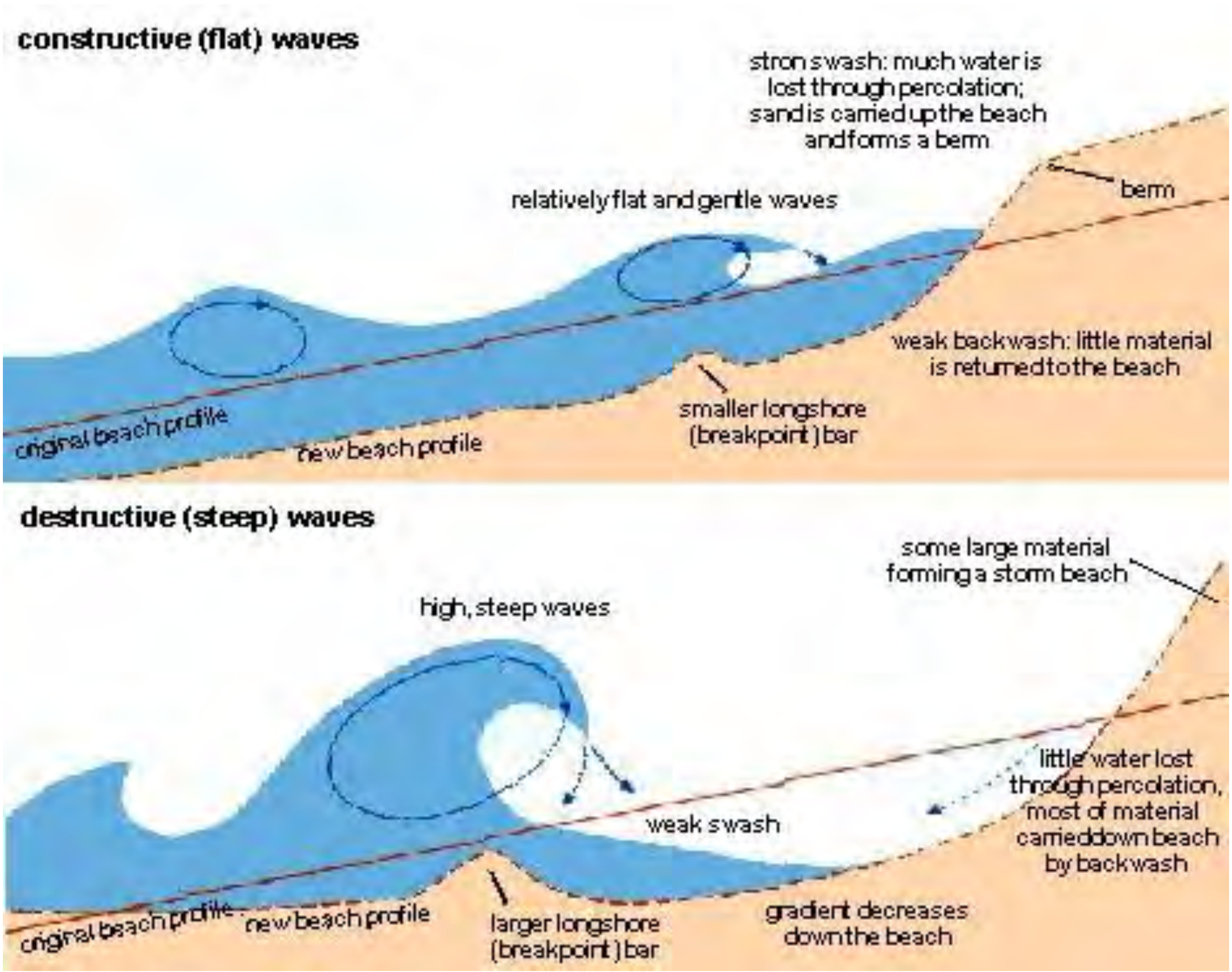


Fig. 24. Constructive and Destructive Waves (by J Harkness).

Constructive waves operate in calm weather and are about a meter high. The swash is strong and erosion is limited. They are involved with the transport and deposition of material creating landforms.

Destructive waves operate in storm conditions and are about 5 to 6 meters high. The backwash is strong and there is a lot of erosion.

The sea erodes the coast in five ways:

1) Hydraulic action – lots of sea water crashes against the land, and air and water are trapped and compressed in rock surface cracks. When the sea moves away again, the air expands explosively weakening the rocks, enlarging the cracks and breaking pieces off.

2) Corrasion is very effective and is caused by broken rock fragments battering the land, cliffs, etc., and breaking off other pieces of rock.

3) Attrition occurs when rock fragments grind each other down into smaller and smoother pebbles, shingle, and finally sand which is later deposited as beaches, etc.

4) Corrosion involves chemical action of sea on rock. If the rocks is limestone, it dissolves in the sea water – some sea salts can also react with certain rocks and cause them to rot.

5) Wave pounding – the “battering ram” action of the weight of the pounding waves.

Waves can move material along the coast:

1) Longshore drift happens when waves break at an oblique angle to the shore (not right angles) due to the prevailing wind.

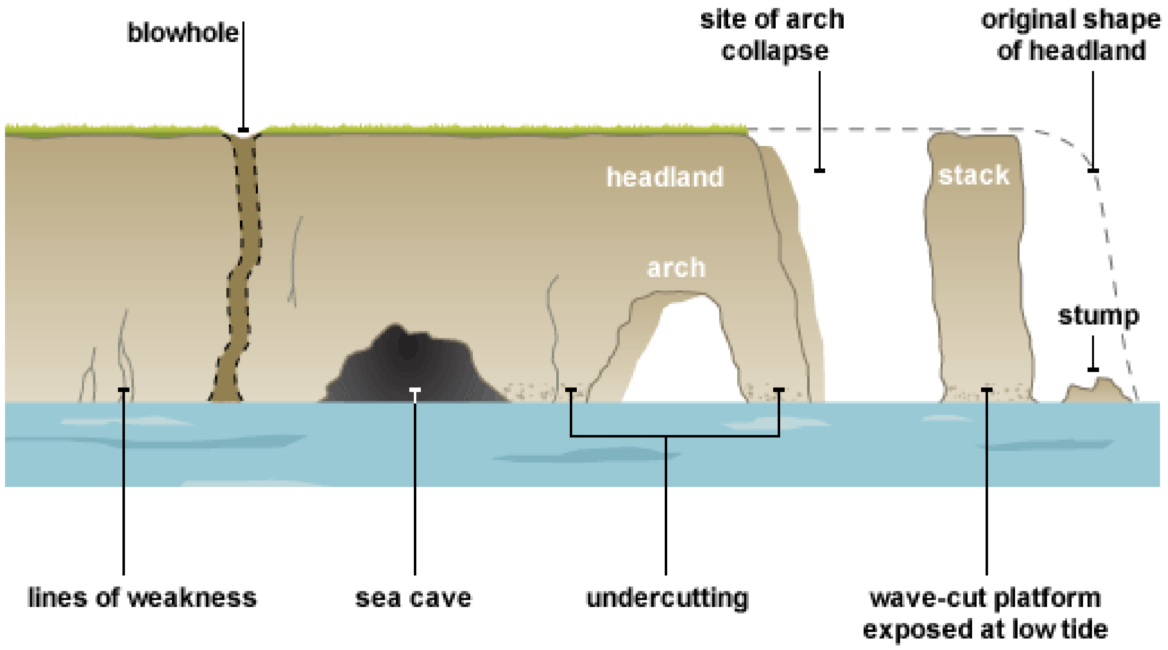
2) This means that each wave pushes material along the beach a bit more.

3) Features can be formed by this process.

Wave erosion forms many coastal features over long periods of time. Rock erosion forms cliffs. Waves erode rocks along the shoreline by hydraulic action, corrosion, corrasion and pounding. A notch is slowly formed at the high water mark which may develop into a cave. Rock above the notch becomes unstable with nothing to support it, and it collapses. The coastline can retreat over many years as this process continues to form a wave cut platform. White Cliffs of Dover may be a good example. The actual size and angle of the cliff will depend on the local rock and its hardness, etc.

Eroded hard and soft rocks form headlands. In Ukrainian they are виступи далеко в море. If there are alternate bands of hard and softer rock in the coastline, the harder rocks take longer to erode than the softer rocks. The hard rock will be left jutting out forming one or more headlands, usually with cliffs. The softer rocks will be eroded to form bays. The bays will usually slope more gently inland, creating room for a beach to form.

Caves, arches, and stacks can also be formed. A crack or rock weakness in a headland can be eroded. This forms one or more caves. Further erosion enlarges the cave and it breaks through the headland forming an arch. The roof of the arch is often unstable and eventually collapses leaving a stack or series of stacks:



(Source: www.bbc.co.uk/bitesize/higher/geography/physical/lithosphere/revision/2/)

Fig. 25. Caves, Arches, and Stacks Formation.

Certain coastal landforms, like the beaches, are formed by deposition. Beaches are found on coastlines where eroded material in the sea has been deposited e.g. in bays between headlands. They vary in size. Storm beaches are ridges of boulders at the landward side of beaches caused by heavy seas piling up material at the high-tide mark. Spits (коси, як Тузла) are long beaches formed by longshore drift. Spits are sand or pebble beaches sticking out to sea, but joined to the land at one end. They tend to be formed by the process of longshore drift. At the spit end there are usually some hooks or reserves formed by occasional strong winds from another direction. Barrier beaches are found where a spit extends right across a shallow bay. The water behind it is left as a lagoon which may slowly become a marsh. As these elements of geology are to be of use to the students of ecology, we shall now speak on ocean life. Living things abound in the oceans. Marine animals are classified b their habits and by the part of ocean water which they use.

Plankton are plants and animals that float at or near the ocean surface. Plankton, for the most part, are microscopic creatures. Diatoms are tiny one-celled plants, a type of algae. The diatoms may cover the ocean surface and form a great blanket of food. Marine animals move in quickly and the food is used up.

Nekton include all swimming forms from tiny herring to huge whales. Nekton can move from one depth and place to another. Some nekton prefer cold water, others like warm regions. Most flesh-eating nekton prefer cold water.

Benthos are bottom dwellers. They live in shallow water where sunlight reaches the seafloor. Benthos include coral, snails, starfish, clams, and other animals that live on the seafloor.

Both oxygen for animals and carbon dioxide for plants are dissolved in seawater. Buoyancy ['bɔiənsi], or upward lift, of water makes movement in ocean water very easy. Plants are the base of the food chain in the ocean as they are on land. Ocean plants can live only as deep as sunlight reaches. Most plants belong to the floating plankton. Scientists have found life even at the bottom of the Marianas Trench which is nearly 11 kms deep.

Sea level is the elevation at which land and sea meet and is defined as zero elevation. Without consideration of tides and waves. All ocean depths are below sea level. Most land elevations are above sea level. In a few places even land elevations are below sea level.

The shoreline is the boundary where land and sea meet. The shore zone includes the area lying between high and low tides. Near the shore, materials are in constant motion. The coast is a strip of land lying between the shoreline and the first big change in the land surface. The width of the coast may be from several meters to several kilometers. It depends on the topography or relief of the land. The coast belongs to the continent, but it is linked to the sea by contact with ocean waves and currents. The continental shelf is a relatively flat part of the continent which is covered by seawater. The shelf lies between the coast and the continental slope. Water from the ocean basin covers the shelf. The continental slope is the steeply sloping surface between the outer edge of the continental shelf and the ocean basin proper. If you think of the ocean basin as a pan, the continental slope becomes the side of the pan (сковорідка).

The abyss [ə'bi:s] is the deepest part of the ocean basin. Its average depth is about 4572 meters. Abyssal plains are almost level areas in the deepest part of the ocean basin. Many volcanic peaks rise from the plain. Single or in chains. Some volcanoes rise above sea level to form islands.

The mid-ocean ridge is the longest feature on the earth. This underwater mountain chain rises from the abyssal plain. Some of its peaks form islands, but most of the chain is covered by water. The ridge is 482 to 4830 kms wide. The mid-ocean ridge extends from Iceland in the Atlantic Ocean to midway between Africa and South America in the South Atlantic. It goes around Africa, into the Indian Ocean, and then across the Pacific northward to the North American continent. The chain is about 75000 kms long. A rift zone, or central crack, is present throughout the length of the chain. This rift zone has many earthquakes, and volcanoes are common.

Trenches are gashes cut into abyssal plain. The Grand Canyon is about 1.6 kms deep. The Marianas Trench, of which it was mentioned here, is at least 11 kms deep. It is in the western Pacific Ocean.

Sediments cover the ocean floor almost everywhere. Deep ocean sediments contain meteoric dust, volcanic ash, some organic matter, and windblown dust.

Ooze [uz] (липкий мул, твань) is the most characteristic deposit of the abyss. Between 4000 & 8000 meters, the deposits are mostly fine clay. Between 600 & 4000 meters, carbonate matter can survive. Shells and diatom coverings cover about half the ocean floor.

Coral reefs are calcite deposits built in shallow water on the edge of the continental shelf. The reef structure is started by coral organisms that live in colonies. Once the reef structure begins, many other forms of life join the colony and add to the reef growth. Conditions favoring such growth are water temperatures between 20 and 26°C. The water needs to circulate freely in order to bring food to the stationary animals. The water also must be shallow enough, so that sunlight reaches the reef. Reefs are common around volcanic islands in the South Pacific. Many of the volcanic islands slowly sink below sea level. The coral reef, however, continues to grow upward towards the sunlight. This kind of coral island is known as an atoll. The ring of coral surrounds quiet water or a lagoon. At some depth below the lagoon the former volcanic island is submerged:

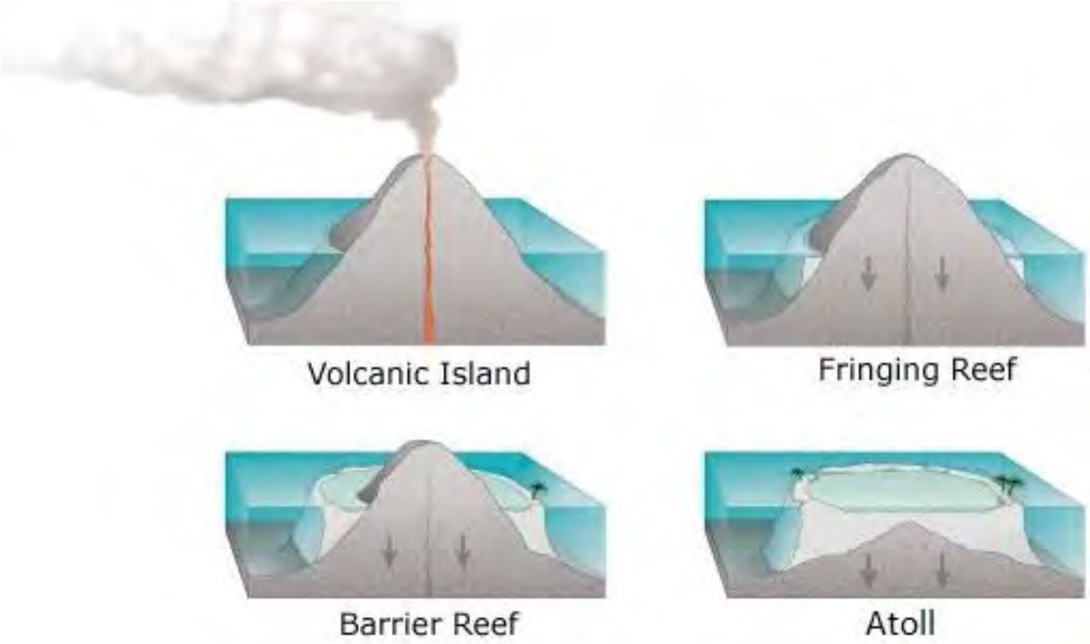


Fig. 26. Deep Sea Deposits (Source: http://pubs.usgs.gov/fs/2002/fs025-02/).

Ooze is red-colored clay in Ukrainian understanding. It may contain concretions of reduced Mn, Fe, S_2, Co, Ni, Cu, Pb, Zn, Mo, V, La, etc.

2.3.6. Geologic Activity of Lakes and Bogs

Freshwater lakes all over the globe contain about $125*10^3$ km^3 of water (Pleiss, 1977). It is only 0.009 and 0.008 percent respectively of total volume of water on Earth. Somebody was quite right in calling the lakes, and ponds, and pools the oases in the landscape. Total area of the lakes is about 2.7 mln km^2 which makes 1.8 percent of the global terrestrial area.

So the lakes are the reservoirs of water on the surface of the continents which are not directly connected with the seas and oceans. The lakes are found at different altitudes, from about -400 to over 5000 meters. Their size and area differ very much from less than a square km to 31 thousand km^2 (Baikal) or even 395 thousand km^2 (Caspian Sea). The depths also differ: Baikal's is of 1741 m, whereas Elton's is only 0.8 m

Lake basins may arise as a result of internal or external dynamics processes. Some are the depressions plowed out by the glaciers and partitioned by moraine ridges. Some lakes formed in tectonic grabens, like Baikal, Issyk-Kool, Victoria, etc. Karst phenomena may also lead to the formation of lakes, as well as the craters of extinct volcanoes. Such lakes are found in Kamchatka or in Armenia. Lakes may also be formed within the flood plains of river valleys (oxbow lakes) and in the deltas.

About 20 thousand lakes, of which over 7 thousand are over 0.1 km^2 in area, are found in Ukraine. More and more men-made lakes (the reservoirs) appear on Earth. Some lakes are locked reservoirs with standing water (Caspian Sea, Aral, Balkhash, Issyk-Kool, etc.). Some are with alternating regime and some – with flowing water (river lakes), like Ladoga, Onega, Baikal, etc. Some, with underground flow (like those formed in karst), are called blind. The lakes may be fed with atmospheric water and with ground water. By the chemical composition of their water, the lakes are fresh-water (water salinity less than 0.1%) saltish (0.1-3.5%), and salty (>3.5%). Flowing-water lakes are, as a rule, fresh-watered especially so in humid climates. Onega's water contains only 0.03% of soluble salts. Standing-water lakes are the most salty.

Geologic activity of the lakes can be classified as destruction, transfer and accumulation of sediment material. Destructive or ruining activity of the lakes is like that of the sea but much smaller. The same is the work of the waves and small tides. Large lakes (Caspian, Aral, Baikal, Onega) have large waves – up to 2-3 meters in height. Their abrasive activity may be very great.

Accumulating or collecting activity of the lakes depends on their water regime, water salinity, size, climate and topography. Mechanical, chemical, and organic sediments form in the lakes.

The sediments of fresh-water lakes have mechanical and organic origin. The sediments of salt ones – a chemical origin. In the zone near the banks – pebbles, gravel, and sand accumulate. Further from the banks, there are found the accumulations of clay and lake silt, or marl. As the lake starts overgrow with vegetation, organic silt – sapropel – forms in the central part of it, whereas near the banks, there form the accumulations of peat. With time, sapropel subsides and becomes a compacted fossil rock, a sapropelite.

The lakes formed by the glaciers have the alternating layers of sandy and clay deposits (bandied clays).

Chemical sediments dominate in salt lakes. Of soluble salts, that crystallize with their drying, the chlorides are the most abundant, whereas the sulfates and, occasionally, carbonates occupy a lesser place. Chemical and organic sediments are raw materials for chemical, food-producing and other industries. Lake muds are employed in medicine. In tropical lakes the deposits of bauxites are no rarity.

Bogs. They are the water-logged areas of land with specific vegetation, after dying out of which peat or sapropel are formed. In Ukraine the bogs are encountered mainly in the Forest Zone (Polissya) and within the floodplains of big rivers. Marshy meadows lie between ponds often overgrown with splendidly colorful plant species. Whatever size, whether pond, pool or lake, all areas of fresh water enliven and enrich the landscape in a variety of ways.

By the conditions of their formation (feeding with water) the bogs are upland (on water shads) and lowland (in river valleys) (Fig. 27).

Upland (ombrotrophic) bogs are fed with only atmospheric water. They are poor in mineral salts so the vegetation is poor and not demanding to nutrition. The residues of vegetation form peats with low ash content and high heat capacity.

Lowland (minerotrophic) bogs are fed with ground water rich in salts and minerals. The vegetation on them is richer. Peats form with high ash content and lower heat capacity as compared with upland peat.

Carbonation of residues in peat formation leads to ever increasing content of carbon in peats, which with time and under the pressure of overlying layers may be transformed very gradually into brown coal.

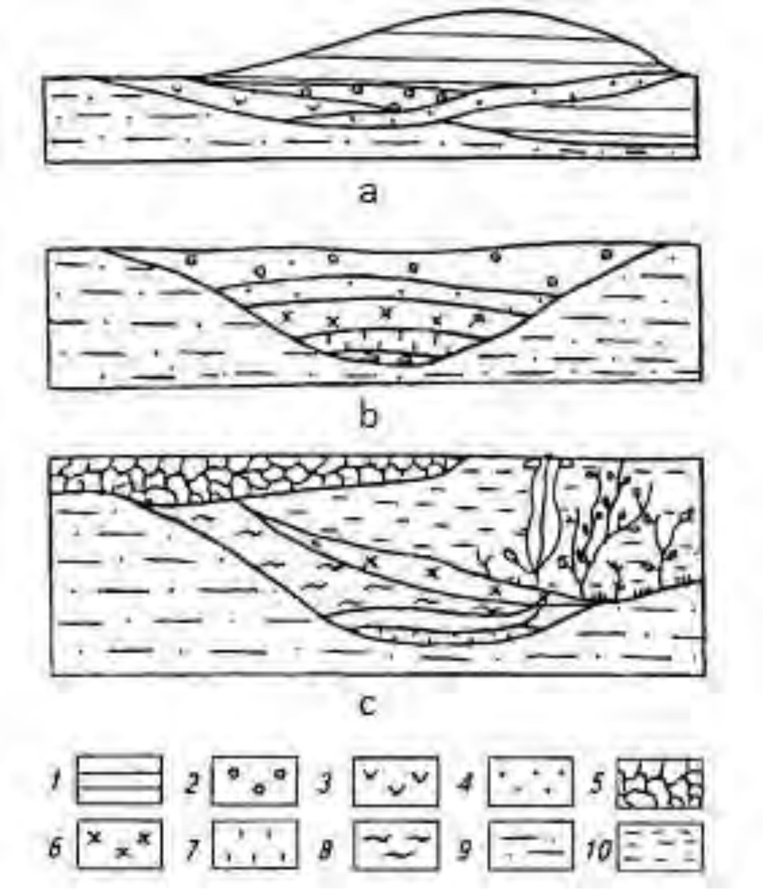


Fig. 27. Genetic Types of Bogs: a – upland (ombrotrophic) bog; b – lowland (minerotrophic) bog; c – swamp forming with the overgrowth of a lake; 1 – sphagnum peat; 2 – sedge peat and sedge-willow peat; 3 – hypnum peat; 4 – reed peat; 5 – floating peat different composition; 6 – sapropel peat; 7 – sapropel; 8 – ooze; 9 – solid; 10 – water

2.3.7. Geologic Activity of Glaciers

A glacier is a large mass of moving ice and snow. Glaciers form where there are many large snowfalls and the temperatures remain very cold. New snow falls in flakes on the surface of the old snow. As the snow builds up, rounded ice granules form in the upper layer of snow. Some glaciers form in high mountains where the snow that falls in the winter does not melt completely in the summer. The snow builds up over the years and gradually turns into ice. When about 30 to 60 meters of snow and ice granules accumulate, their weight causes bottom granules to become crystallized ice. The bottom ice becomes pliable, somewhat like putty (замазка), and begins to move. If there is a slope, movement will be downslope. If there is not a slope, the glaciers spred outward in all directions, like spilled molasses (патока).

Glacier classification. Glaciers are classified as valley, piedmont ['pi:dmɔnt], and continental. Valley glaciers are also called alpine glaciers. They form at high elevations where snow lasts year after year. Valley glaciers take over the valleys of a river system that preceded the cold climate. Glaciers begin as snowfields that collect in hollows, or valleys. Steep slopes have little snow because the snow slides down the slope, is blown away, or melts. But in the hollows the snow can remain.

Changes in climate bring about changes in the elevation of the end position. Warmer climates cause the glacier to retreat to a higher elevation. Colder climates move the glacier front to a lower elevation.

Continental glaciers are great masses of ice that accumulate in high altitudes. At present, Greenland and Antarctica are covered by continental glaciers. Movement of ice in these glaciers is outward from the center of accumulation. Continental glaciers range in thickness from 2500 to 3000 meters.

Piedmont glaciers form at the foot of some mountains where valley glaciers extend onto the plains. At the foot of the mountains, ice forms a continuous sheet as the glaciers spread out on the plains and join one another.

Erosion by glaciers. They carry a more massive load of debris that rivers. But they do not cover as much of the earth. However glaciers, where they are present, completely change the face of the earth. They move over the land like giant bulldozers. They scrape away all loose fragments and pile the debris in massive mounds. Glaciers work by abrading and plucking. Abrasion is a process where rock fragments in the ice grind against the rock over which the ice is moving – like a rough sandpaper – wearing away the land. Plucking is the combination of freezing with the ice flows downward until it penetrates openings in the rock. Usually, those openings are cracks. Meltwater freezes in the cracks and becomes part of the moving glacier. In this way massive loads of boulders, gravel, and sand are added to the bottom of the glacier. As the ice moves forward, the debris abrades the bedrock beneath. Rocks beneath the glacier may be scoured or polished. Glaciers dig into soft rock more deeply than into hard rock.

Glacial erosion created upland features. Ice in hollows caused plucking and freeze – thaw action which steepened the back and side walls. The ice moved with a circular motion known as rotational slip which deepened the hollow into a bowl shape, a corrie – forming a lip at the valley end. The original hollow, or cirque, in which the snow accumulates may eventually reach the mountain summit. Hollow in Ukrainian means “лощина”. Some peaks look like horns (роги) after it. Two neighboring corries meant glacial erosion narrowed and steepened the wall between them, forming a knife-edged ridge or arête.

The old valleys were altered by the glacier:

1) The valley cross-profile was changed from a V-shape to U-shape as the glacier moved down it with great power, eroding slope material.

2) The valley, or glacial trough, became a straighter, more linear features, as the ice didn't wind round spurs but cut through them, leaving steep edges on the valley sides – truncated spurs.

3) Original tributary valleys are now at higher levels than the main valley as they didn't experience such powerful glacial erosion. Tributary streams enter the valley as waterfalls from hanging valleys.

When the ice melts, a variety of deposits are laid down – either by the glaciers themselves or by powerful meltwater streams. Glaciers produced moraines, drumlins, and erratics.

At some time in the life of a glacier, melting begins. Then the ice drops its load and forms a number of new landforms. The first material dropped as the ice melts is unsorted and unlayered. It is a jumble of boulders, sand, and clay called till or drift. As meltwater flows away from the front of the glacier, it carries some of the sand and clay with it. Like river deposits, this sediment is sorted as the velocity of the water slows.

Most sediment is dropped at the front of the glacier. Mounds of till form during periods when the forward movement of ice is about equal to the rate of melting. These mounds or moraines, cover the front of a glacier. There are different types of moraines. Till deposited at the front end of a glacier is called a terminal moraine. Till deposited along the sides of a glacier is called a lateral moraine. The rocks found in a moraine are evidence of where the glacier formed. Rocks can be carried great distance. The position of a terminal moraine indicates how far the glacier advanced before retreating.

Drumlins. A drumlin is an oval-shaped mound of till. Its tip points in the direction that the glacier was moving. Drumlins were formed as deposits of till rounded by the glacial ice.

Meltwater deposits. Melting ice forms streams that flow out from the glacier, the meltwater streams. They carry away sand and gravel. The meltwater may also form small lakes and ponds near the glacier. Many present-day rivers were originally meltwater streams.

Sediments deposited by rivers of glacial meltwater form areas called outwash plains. They are fan-shaped and form in front of terminal moraines.

Iceberg deposits. Valley glaciers and continental glaciers sometimes reach the sea. When this happens, the glaciers form cliffs of ice and snow. Parts of the glaciers break off and drift into the sea. These glacial parts are called icebergs. Icebergs may contain rocks and debris picked up from the land. As the icebergs melt, the rocks and debris are deposited on the ocean floor.

Glacial lakes. Glaciers created many of the lakes in the United States. In Finland too. Such lakes are found in Ukrainian Forest Zone too. Glaciers can form lakes in two ways. Glaciers till or deposits from meltwater sometimes pile up in low-lying river channels and other areas. These deposits keep water from flowing away from the area. The land areas fill with water, and lakes are formed.

Sometimes huge blocks of glacial ice are left behind by a glacier. The ice blocks are surrounded by or covered with sediments deposited by the glacier. When the ice melts, it leaves a depression, or hole, in the ground. Lakes formed in this way are called kettle lakes. Kettle lakes are usually round and very deep.

Other meltwater deposits. Eskers are long, winding ridges of deposited material laid down in the channels of streams in and under ice. Kames are small mounds of deposits formed when meltwater streams flow over the snout (рило) of the glacier to flatter ground below. Kame terraces are linear features along the side of valleys where meltwater streams deposited their load between the side of the valley and the edge of the glacier.

Climate and Sea Level Changes. Global climates were affected by the melting of the glaciers. Even regions not a part of glaciations became more humid. Many huge lakes formed as the glacier melted. As the ice melted and returned to the ocean, sea level rose. If all the ice left in Greenland and the Antarctic melted, sea level would rise about 60 meters.

Glacier velocity is from a few dozen centimeters to 20 meters per day.

Theories of Glacial Origin. Geologists are certain that ice ages have occurred at least 3 times in the past. Probably there have been other glacial periods, but these 3 ages left confirming evidence. The earliest ice age occurred over 600 million years ago. The second period of glaciations took place about 200 million years ago. The most recent ice age began over 1 million years ago, and the ice retreated to its present position about 10000 years ago. The most recent research suggests that the amount of ice on the earth is greatest every 22 000, 42 000, and 100 000 years. Scientists relate these figures to three changes in the earth's movements. Earth's axis is now tilted 23½° to the plane of its orbit. This tilt increases to 25°C and then decreases to 22°C. When the tilt is at its maximum, 25°, Earth's poles receive more radiation. When the tilt is at its minimum, 22°, the poles receive less radiation. This decrease occurs about every 42 000 years. Another factor may be the change in the shape of Earth’s orbit around the sun. Every 100 000 years, the orbit becomes almost a circle. When the orbit is more circular, than at present, Earth is farther from the sun during fall, winter and spring. These seasons then are colder than they are now.

The third factor also involves the earth’s axis. Due to differences in gravity force on the earth, the axis wobbles every 22 000 years. According to this theory, an ice age may begin. The more elliptical orbit and greater tilt bring warm interglacial periods. Changes in the wobble may make the ice age colder than usual and the interglacial periods warmer.

For the past million years, ice ages have alternated with warm climates. Some of the ice ages lasted 60 000 years. Some interglacial intervals were warmer than our climate today. The most recent ice age reached a peak 18 000 years ago, then the earth began to warm up. The ice retreated to its present location about 10 000 years ago. The present interglacial period began about 14 000 years ago Most of the interglacial intervals have lasted about 12 000 years. Some scientists predict the beginning of another ice age within a 1000 years or so. Present theories do not answer all questions about glaciations, but perhaps current research will help.

2.3.8. Geologic Activity of Ground Water

Ground water is an important part of the water cycle. About 0.25 cm of the average 76.2 centimeters of rain that falls each year returns to the underground. Underground water returns to the surface in permanent streams, swamps, springs, and seeps. Where the ground water is removed through wells, it may be used up faster than it can be returned by rainwater.

Underground water moves downward through permeable rock until it reaches a depth where no rock openings are present. Above the impermeable rock, all openings are filled with water to form a zone of saturation. The upper surface of the zone is called the water table. Above the zone of saturation, rock openings are filled with air between rainstorms. In this zone, called the zone of oxidation, water and oxygen react with elements in the rock to form clay and carbonates. This is the zone where the rock is changed to soil. Some water that sinks into the ground becomes soil moisture for a time. This zone, although not too thick, supplies water to plants. Therefore, it is important. If soil moisture is not renewed by rainfall, it is lost to the underground water or to evaporation. If present, soil moisture is just below the ground surface.

Some water is trapped just above the water table in small openings called 'capillaries. These hairlike openings attract molecules of a liquid (surface tension) and return water to the surface during dry spells. Now it is important to understand that soil moisture is measured in percent by dry weight. In any layer of soil, having the same moisture all over, it is possible to calculate the amount of water in metric tons or cubic meters per hectare by the formula:

$$Q = W * h * d, \tag{2}$$

where Q is the amount of water in a layer h centimeters thick and having a bulk density in grams of dry soil per cubic centimeter. Bulk density is the weight in grams of cm^3 of dry soil in its natural (undisturbed, like it lies in the field) compactness. Having divided Q in mt/ha by 10 we can obtain the amount of water in a given layer expressed in millimeters of water “column”. It is convenient to express the amount of rainfall and the amount of water in a certain layer of soil in the same units like the millimeters.

Water balance in a root zone of a plant is depicted in the following scheme:

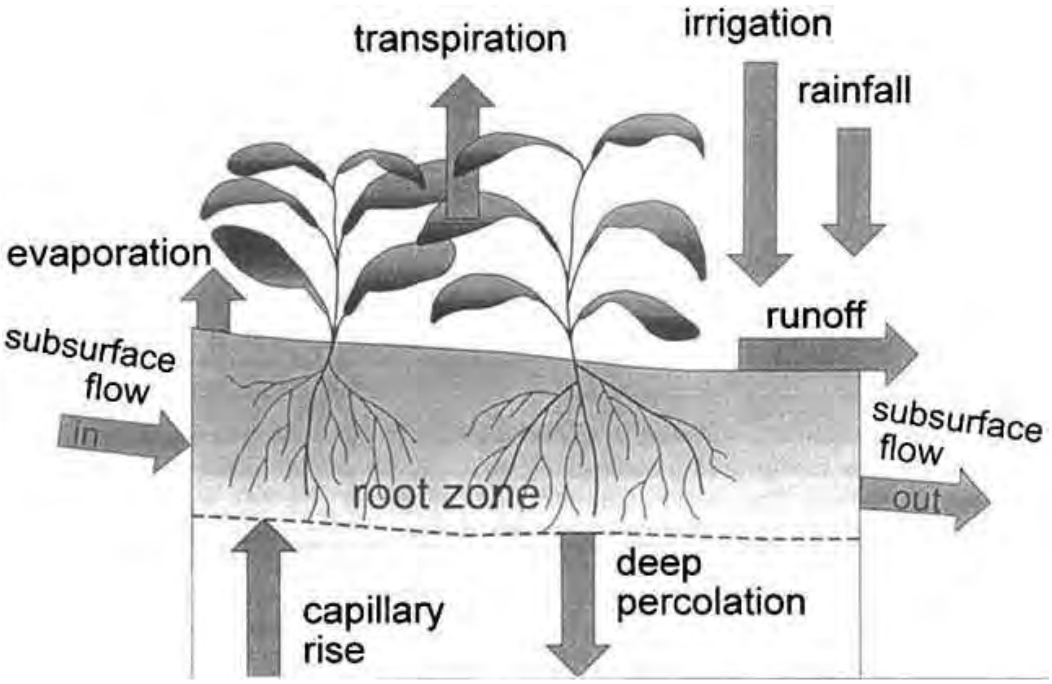


Fig. 28. The Water Balance of a Root Zone (Schematic) (D. Hillel, 1982).

Rainwater moves directly down to the zone of saturation if all rocks are permeable. If some layers are impermeable, without connected openings, water may be trapped above them in a perched water table. Then water flows along the surface of the impermeable layer until it reaches the opening of some kind.

Gravity causes ground water to move from higher to lower elevations. Ground water moves more slowly than surface water because of friction with rock openings. Large openings cause less friction than small openings.

Soil water (вода в грунті) is found in soil layer. It can get warmer in summer, evaporate from the soil, and freeze to ice in winter. It can be saline in warm climates, rich in organics in wet (overwet) soils, etc. Ground waters (підґрунтові води) lie at a certain depth from ground surface. They are underlain by the first from the surface impermeable layer. They feed “at the expense” of infiltration and condensation. River, lake, reservoir and other waters may take part in their “feeding”. They do not freeze in winter. Their temperature changes only in a narrow range. Their upper surface (water table) imitates the land forms above them but with a lesser range.

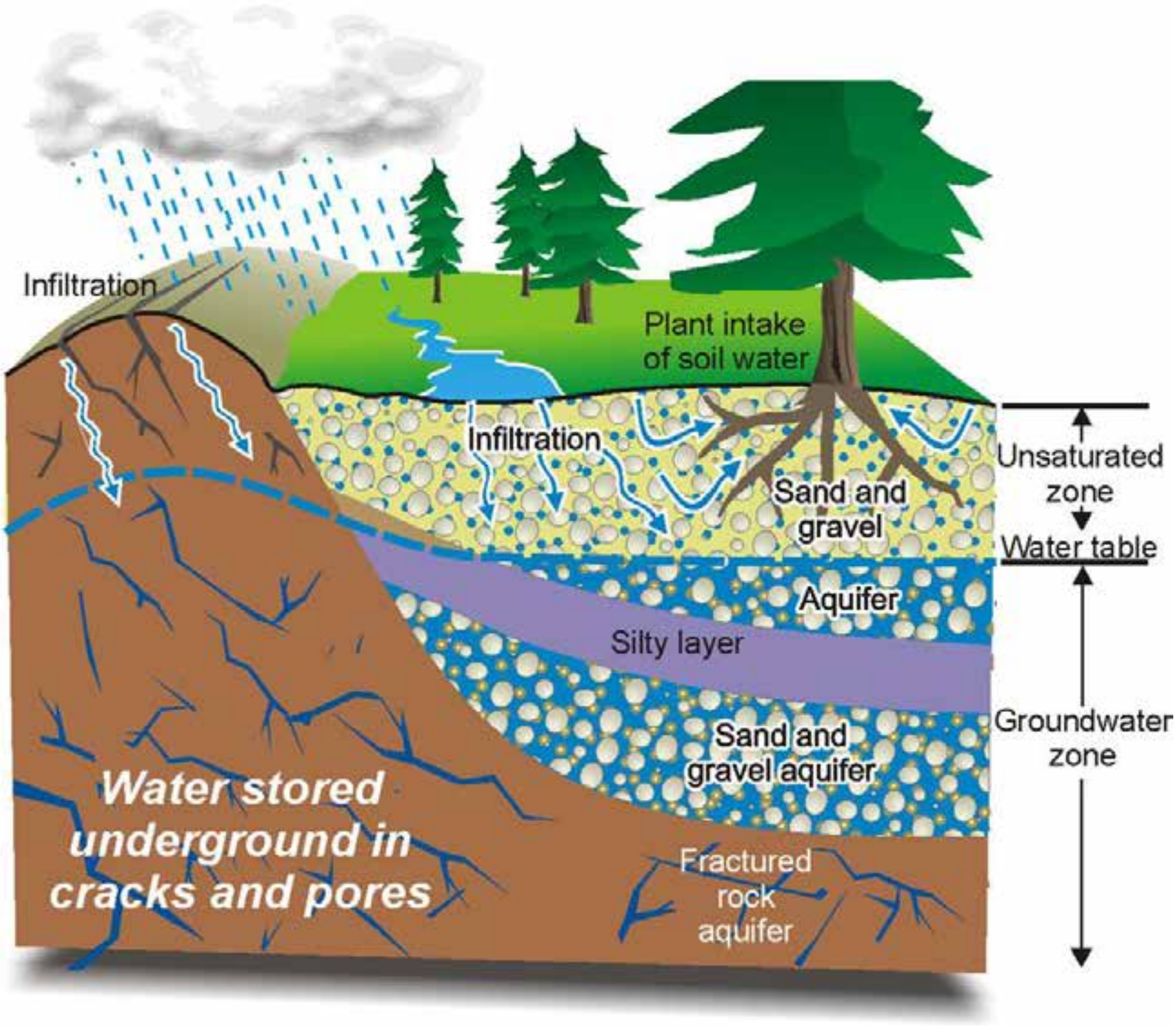


Fig. 29. Ground Waters.

Underground waters (підземні води) lie in the acquifers below ground water.

Permeable rock layers filled with water are called aquifers. Gravel and sand are the best aquifers because they have the largest continuous openings. Openings in limestone along cracks or between layers also may be large. Sometimes all openings have been filled with cement and rocks cannot hold liquid. Most sands and gravels yield water when wells are drilled into them. This water must be pumped to reach the surface.

Artesian water is under pressure due to the weight of the column of water in the aquifer. When an aquifer is between two impermeable layers and the upper part of the aquifer reaches the surface, the water moves through the aquifer as it does in a pipe. Water rises toward the surface through any available opening which may be cracks in the overlying rock layer or drilled wells. Water does not rise as high as the intake area because of friction.

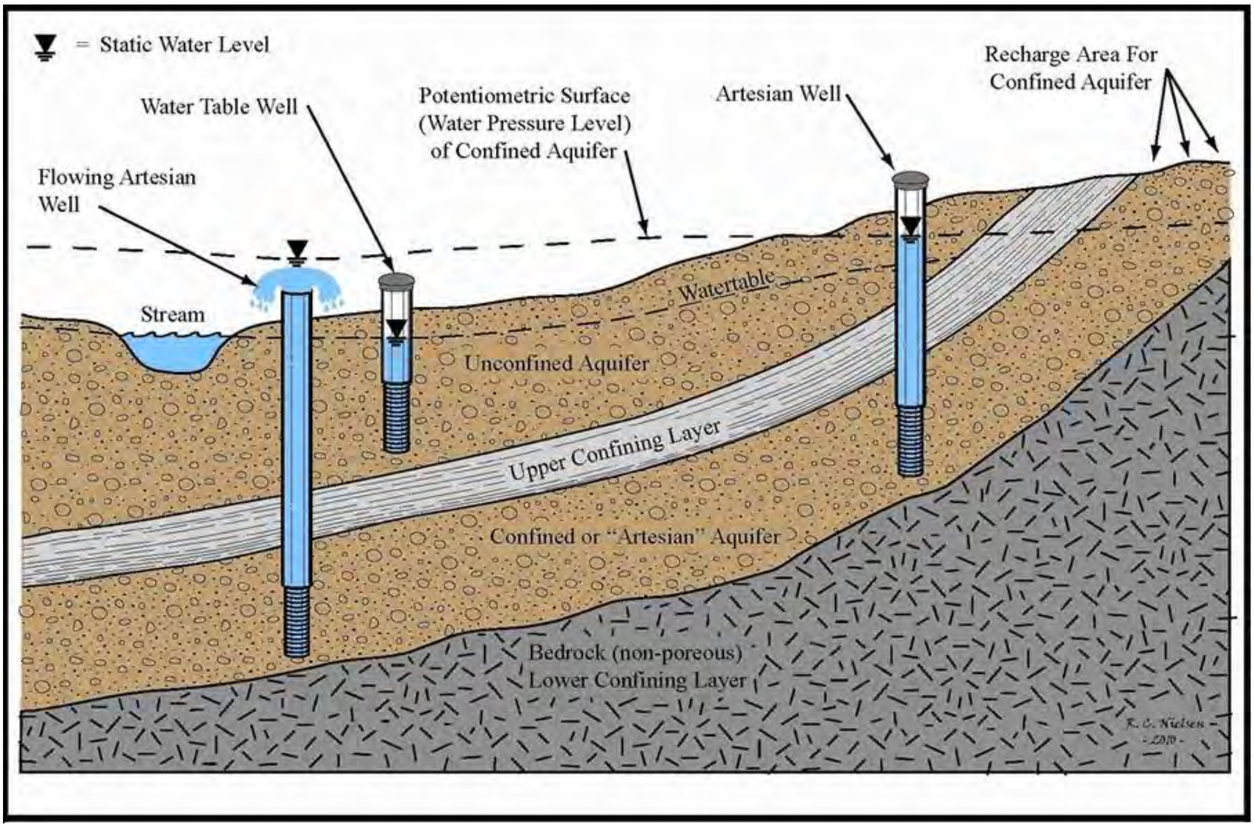


Fig. 30. A cross section of a landscape showing an aquifer carrying water to an artesian well (R. C. Nielsen, 2010).

Geysers are hot springs that have a small surface opening. Water is forced upward by stream pressure at fairly regular intervals. Hot springs occur in the same area as geysers, but openings are not constricted. Calcium sulfate, calcium carbonate, and various sulfur compounds are dissolved in the hot water giving such springs unusual color, taste, and odor. Heat comes from hot, intrusive igneous rocks.

Erosion by groundwater. Rainwater is a weak acid which will dissolve limestone rocks. Water circulating in limestone areas is called "hard" because it contains much Ca^{2+} and Mg^{2+} in solution. As water seeps downward in limestone regions, the limestone is dissolved. This leaves openings called caves. Large caves or caverns may be joined by narrow passages through which a stream of underground water flows. But eventually, the stream rejoins the surface drainage system.

Features formed by solution include natural bridges, sinkholes, and cave deposits. Valleys form when the roof of an underground passage (Karst) collapses. If a section of the roof remains, it forms a natural bridge. Sinkholes (Карстові лійки) are funnel-shaped depressions dissolved from limestone along intersecting cracks or joints.

Groundwater deposits. Groundwater usually contains large amounts of dissolved minerals, especially if water flows through limestone or through volcanic rocks. Deposits from groundwater usually are $CaCO_3$ or silica. Cave deposits formed from cool water are mostly $CaCO_3$. Groundwater loses CO_2 as it seeps into the roof of a cave. Water loaded with dissolved calcite drips from the roof leaving behind a deposit called a stalactite.

Pronounce it [stuhLAKtite]. The rest of the water falls to the cave floor and evaporates, building a stalagmite Pronounce it [stuhLAGmite]. These two forms often merge to form columns.

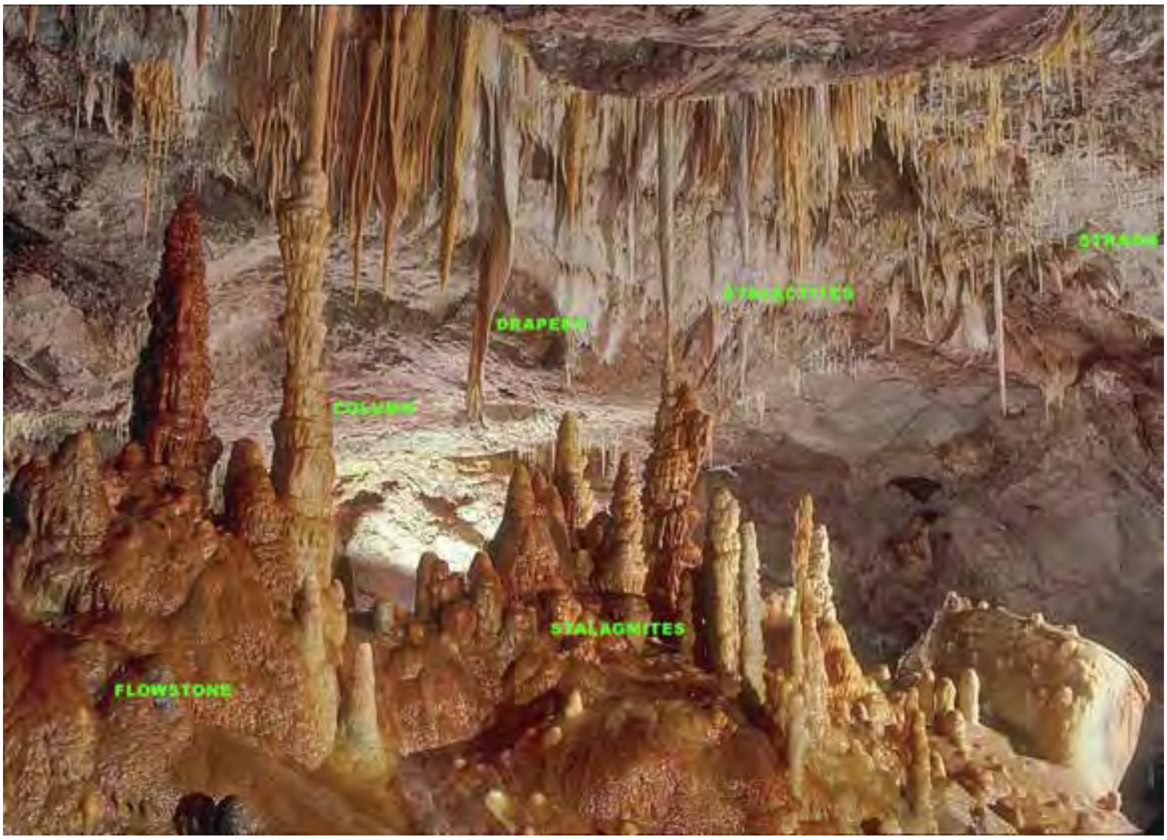


Fig. 31. Stalactites and Stalagmites (from Wikipedia).

Valuable minerals including copper, silver, lead, and zinc are deposited by groundwater in rock openings called veins. Groundwater also deposits cementing material and changes loose sediment into hard sedimentary rock. But of that more anon.

2.3.9. Sediment Diagenesis

The sediments, formed as a result of geological work described in the preceding sections, gradually transform into sedimentary rocks. This transformation proceeds under the influence of physical, chemical, and biological processes. It is known under the name of diagenesis (from Greek = rebirth).

The phenomena of diagenesis envelop the processes of compaction, cementation, recrystallization, dissolution and replacement of minerals; the formation of other minerals as well as of concretions and other forms of mineral existence. The extent and character of mineral transformation depend on sediment texture and composition. Some sediments of sandy texture, or gravels, change little. Some undergo deep changes. Some complex chemical reactions take place in clay and silt sediments, like dissolution and removal of certain mineral components as well as formation of new minerals. Some minerals change their crystal structure. Bacteria and other microorganisms take an active part in the processes of diagenesis.

The processes of recrystallization sometimes occur rather quickly. Organic essence of coral reefs quickly transforms into limestone. The same is true of limestone algae. The rapidness of the process is caused by the evolution of carbon dioxide as a result of organic matter decomposition. The same is the cause of rapid transformation of carbonatic silts into limestone.

The changes of reducing conditions into oxidizing ones and vice versa also favor the rapid character of chemical transformation. When one sediment becomes quickly covered by the other, all oxidized forms of elements undergo reduction. Hydrous micas, containing iron, transform into glanconite under slightly oxidizing conditions, into ciderite under slightly reducing conditions, and into a pyrite. Microorganisms help these processes to proceed fast. They consume oxygen and respire carbon dioxide. Dissolution and recrystallization of rocks favor the formation of concretions and different crystal combination in porous rocks. Cementation of sediment components during diagenesis may occur at the expense of sediment materials or from materials which form at the later stages of sediment transformation (epigenetic cement), e.g. iron oxides, opal, and carbonates.

The processes of diagenesis are “to blame” for the formation of oil, gas and other fossil fuels during the transformation of organic remains. The transformation of plant residues in anaerobic conditions into peat and subsequently into brown coal also deserves the name of diagenesis.

The particles of silt, transferred by the wind to the territory of preset-day Ukraine, underwent compaction, carbonation and organism impact before they became a loess, an ideal parent material of Ukrainian soils.

Questions and Assignments

1. Which geological processes are called endogenic, and which – exogenic?
2. Characterize the essential of endogenic processes.
3. Which types of weathering do you know?
4. Define cortex of weathering and give a list of its basic types.
5. Outline the main aspects of geological work of the wind.
6. Define river system and name its component parts
7. Characterize floodplain deposits.
8. Delta deposits as interaction of geological work of rivers and the sea.
9. How do peats form in the evolution of lakes?
10. Say what you know on the periodicity of glaciations on Earth.
11. Speak on landforms created by the glaciers.
12. Illustrate the difference among glacial till, moraine, and glacial outwash.
13. Speak on groundwater geological activity.
14. Speak on the role of groundwater in formation of soils.
15. Define diagenesis and briefly characterize its component processes.

Chapter 3

Fundamentals of Mineralogy

3.1. Detinition of a mineral

The geologist is primarily interested in the earth's rocky crust, while the soil scientist is interested in sedimentary and to a lesser extent in the products of weathering of igneous and metamorphic rocks, as parent materials of soils. But before we can study rocks, it is necessary to know something about minerals, for these are the building blocks of rocks, and, therefore, of the earth's crust as a whole. The geologists differ when defining the term mineral, but the following definition is generally accepted: Minerals are chemical elements or compounds which occur naturally within the crust of the earth. Some definitions stress that minerals are both physically and chemically determinate natural bodies formed within the earth's crust and on its surface as a result of physicochemical processes. Anyway, the word "mineral" is derived from the Latin "minera" which means "ore". Over 2500 minerals have been found in nature. But only some 50-60 of them compose rocks and are widespread in nature. They are called the rock-forming minerals. Such minerals as calcite, quartz, feldspar, micas, amphiboles, pyroxenes, and many others are rock-forming minerals. The minerals like gold, diamond, uranium minerals and silver are found in relatively few rocks.

Besides a definite chemical composition or range of composition, the minerals possess an orderly internal arrangement of atoms (crystalline structure), and certain other distinct physical properties. It should be noted, however, that the chemical and physical properties of some minerals may vary within definite limits. Rocks are aggregates or mixtures of minerals, the composition of which may vary greatly. Granite typically contains three minerals – feldspar, mica, and quartz, while limestone is composed primarily of one mineral – calcite.

According to their origin, the minerals are divided into primary and secondary. Primary minerals are those that have solidified from an original molten state (from magma) or were formed during the metamorphism. They did not undergo any chemical transformations since their formation. Secondary minerals were formed as a result of chemical weathering of primary minerals. The most widespread primary minerals in rocks and soils are feldspars, quartz, amphiboles, pyroxenes, and micas. These minerals make up the lion's share of the mass of igneous rocks (by F. I. Clark) (Table 7):

Primary minerals	%
Feldspars	59.5
Quartz	12
Amphiboles (hornblendes) and pyroxenes	16.8
Micas	3.8
Others	7.9

Quartz is the most resistant among them to chemical weathering. It is, therefore, the most widespread in soils (>40%), while feldspars occupy the second place (up to 20% in most soils and corti of weathering).

Among the secondary minerals, the most wide-spread are the salts, oxides and hydroxides and clay minerals, such as calcite ($CaCO_3$), gypsum ($CaSO_4*2H_2O$), gibbsite ($Al(OH)_3$), hematite (Fe_2O_3), goethite (FeOOH), kaolinite ($Al_4(OH)_8[Si_4O_{10}]$), etc.

In the majority of soils (except lateritic ones), the primary minerals predominate. Only in the fine separates of soil mechanical particles (<0.001 mm) the secondary minerals predominate.

3.2. Elements of Crystallography

When crystalline minerals, which predominate in nature, solidify and grow without interference, they will normally adopt smooth angular shapes known as crystals. The planes that form the outside of the crystals are known as faces. These are related directly to the internal atomic structure of the mineral, and the size of the faces is dependent upon the frequency of the atoms in the different planes. The shape of the crystals and the angles between related sets of crystal faces are important in mineral identification.

Crystals of all minerals are conceptualized in terms of an array of points in space, called a lattice. The lattice represents the periodic nature of the mineral structure regardless of the specific atoms involved. The unit cell is the repeat unit by which the crystal lattice can be built. The lattices can be atomic (diamond, sulphur, graphite), ionic (halite, gypsum, calcite, fluorite, pyrite, clay minerals, etc.) and molecular (sugar, aspirin, etc.).

All crystalline substances are anisotropic: their properties are different in nonparallel directions. Each mineral has been assigned to one of six crystal systems. These systems have been established on the basis of the number, position, and relative lengths of the crystal axes – imaginary lines extending through the center of the crystal (Fig. 32):

from http://www4.nau.edu/meteorite/meteorite/book-glossaryc.html:

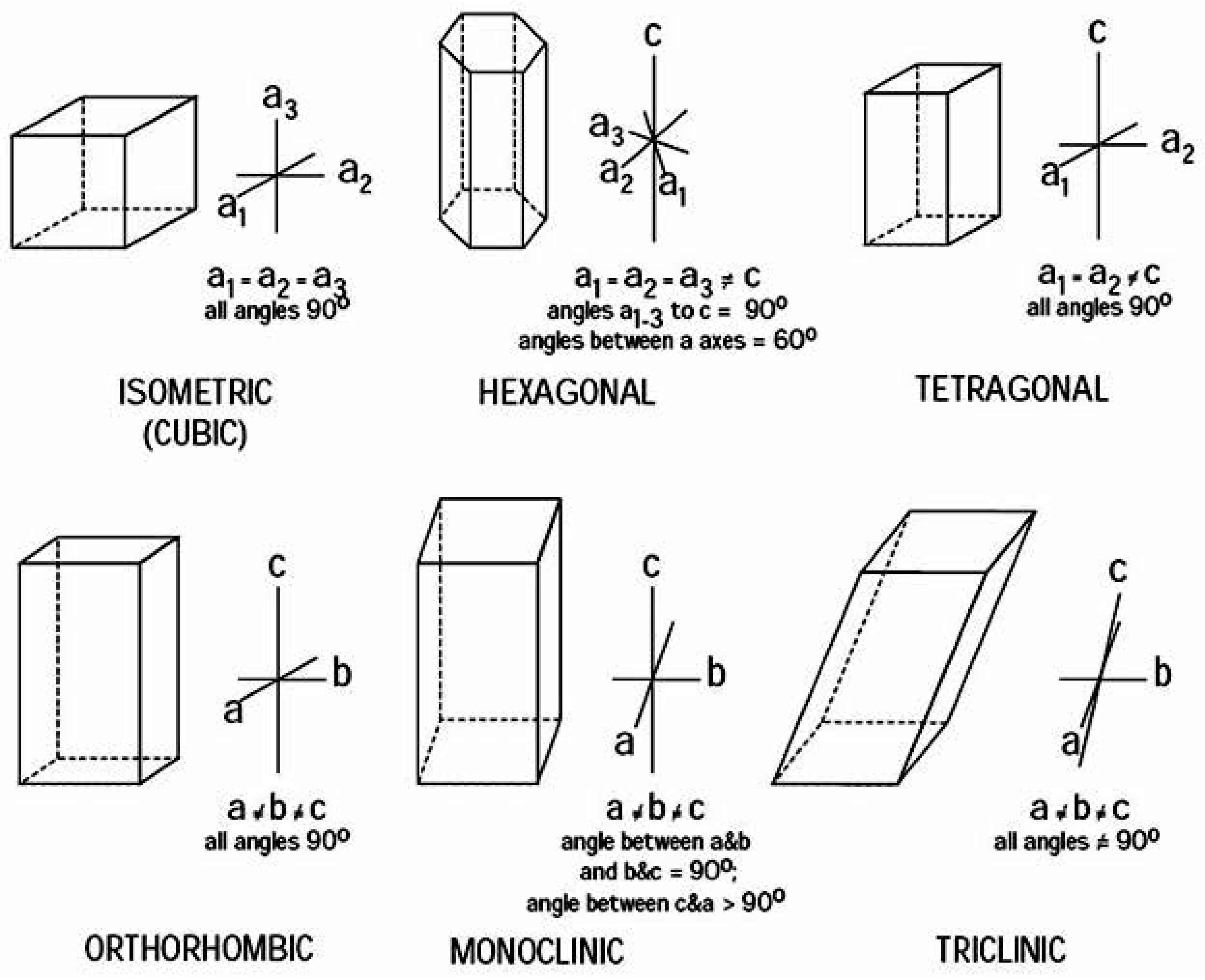


Fig. 32. Crystal axes (shown as lines) and systems.

Crystal axes are used to orient the crystal when it is studied. For example, the crystals, assigned to the tetragonal system, have three axes: two of equal length called the horizontal axes; the third (vertical) axis, which may be longer or shorter than the other two, is always placed vertically when the crystal is properly oriented. Thus, the vertical axis of a crystal will always be perpendicular to the horizontal axes when the crystal is properly oriented.

Every crystal system possesses a type of symmetry that is peculiar to all members of any given system but unlike that of crystals belonging to other systems. The type of symmetry present is determined by arrangement of the axes of the crystal. Mineralogists recognize the following crystal systems:

Isometric of Cubic System. Crystals have three axes of equal length and at right angles to one another.

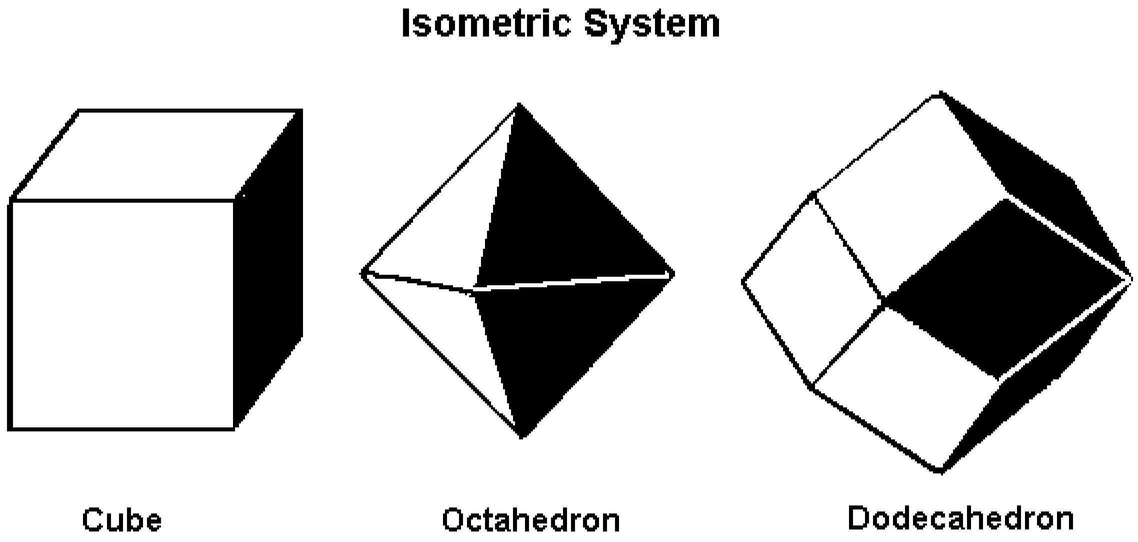


Fig. 33. Isometric Crystal System (halite) (the axes are shown as broken lines)

Tetragonal System. Tetragonal crystals are characterized by having all three axes at right angles. The lateral (horizontal) axes are of equal length but are longer or shorter than the vertical axis.

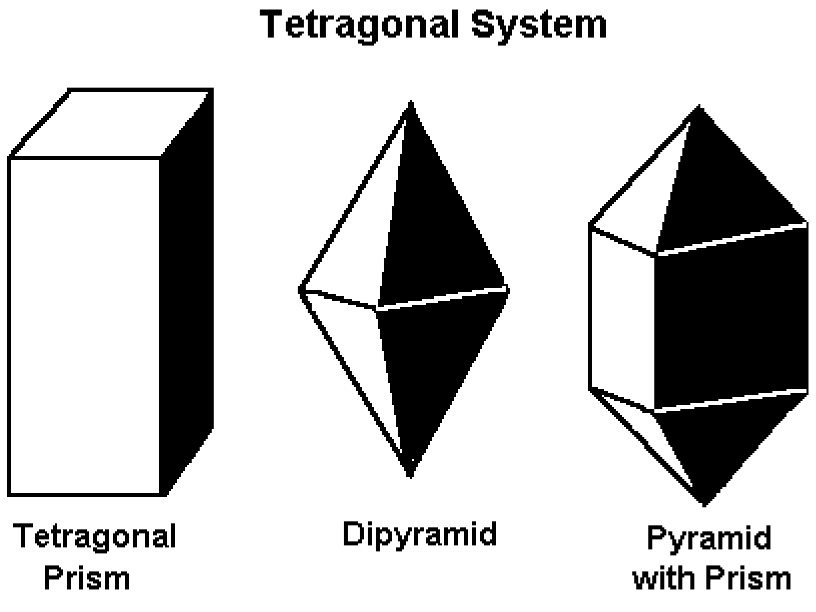


Fig. 34. Tetragonal Crystal System (zircon).

Hexagonal System. This system has crystals marked by three horizontal axes of equal length which intersect at angles of 120 degrees, and a vertical axis at right angles to these. The vertical axis is either longer or shorter than the horizontal axes.

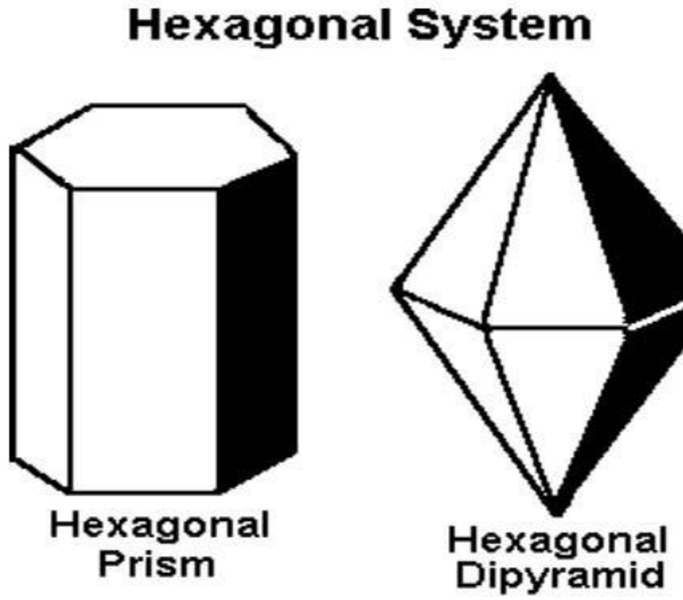


Fig. 35. Hexagonal Crystal System (quartz).

Trigonal system. The trigonal system is a subsystem of the hexagonal. Most gem references will list these as hexagonal. Minerals that form in the trigonal system include all varieties of Quartz, (agates, aventurine, chalcedony, jasper, tiger eye, citrine, amethyst, etc.,) Benitoite, Corundum, (rubies and sapphires,) Hematite, Rhodochrosite and Tourmaline.

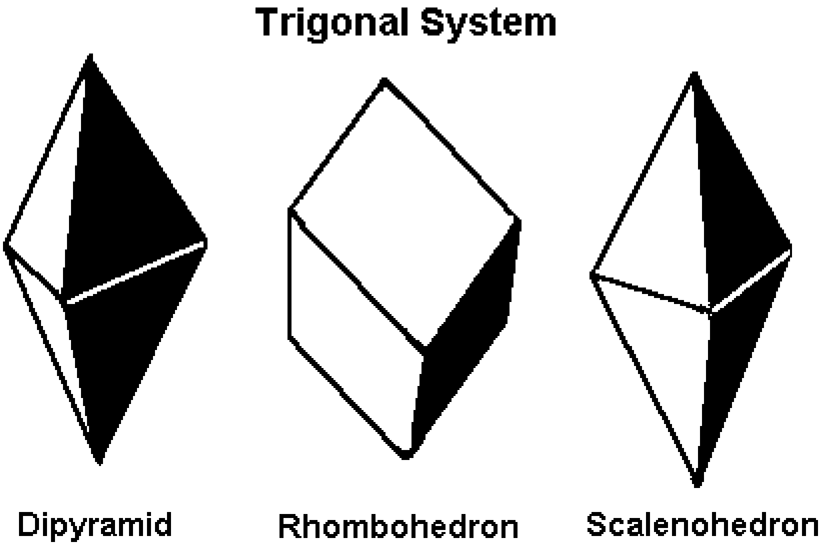


Fig. 36. Trigonal Crystal System.

Orthorhombic System. Crystals have three axes, all at right angles and each of a different length.

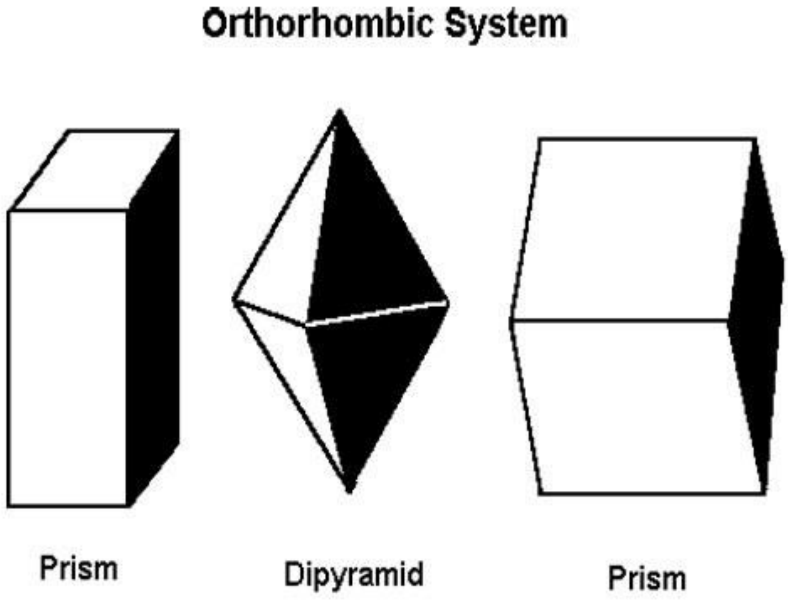


Fig. 37. Orthorhombic Crystal System (staurolite).

Monoclinic System. Crystals have three unequal axes, two of which intersect at right angles. The third axis is oblique to one of the others.

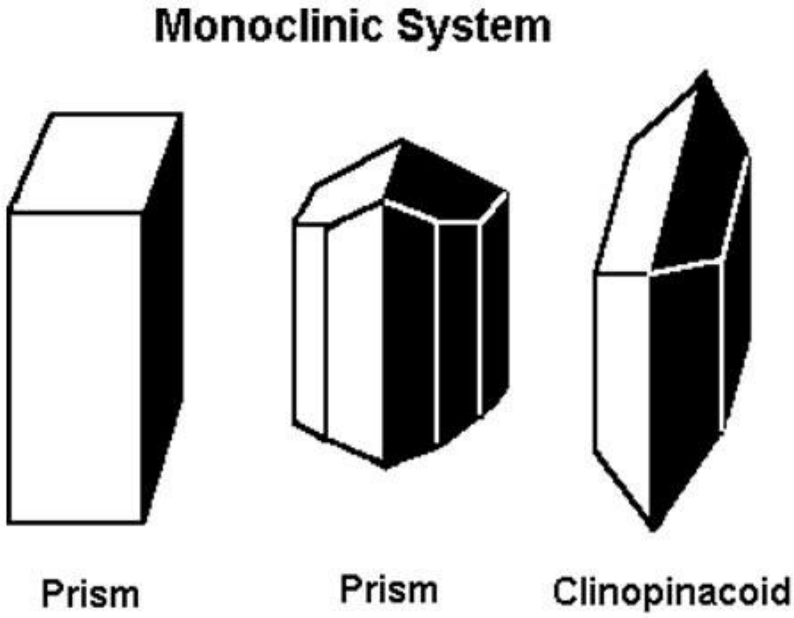


Fig. 38. Monoclinic Crystal System (orthoclase).

Triclinic System. The triclinic system has crystals characterized by three axes of unequal length and all oblique to one another.

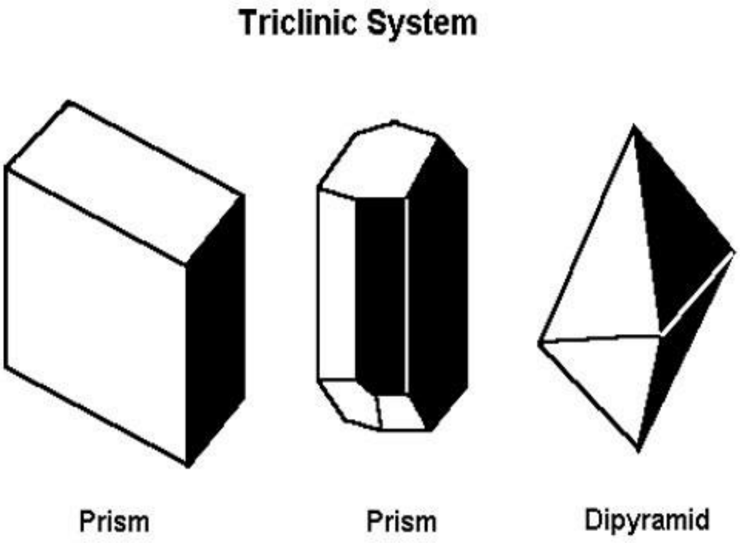


Fig. 39. Triclinic Crystal System (albite).

Crystal Habits. Any given mineral crystal will grow or develop in such a way as to form certain typical shapes. These forms are called their habits. Crystal habits are useful in the identification of minerals because they indicate what forms or combination of forms a mineral is likely to assume. For example, tourmaline has a columnar habit, galena a cubic habit and barite a tabular habit. Because crystals of a specific mineral will develop only in the crystal system of that mineral, it follows that a cubic system crystal will show only cubic system characteristics. However, when crystals are formed at different temperatures, they may assume different habits within their system. Thus, fluorite crystals formed at low temperatures have a cubic form while those formed at high temperatures have an octahedral form. In some instances a mineral crystal may show a combination of two forms.

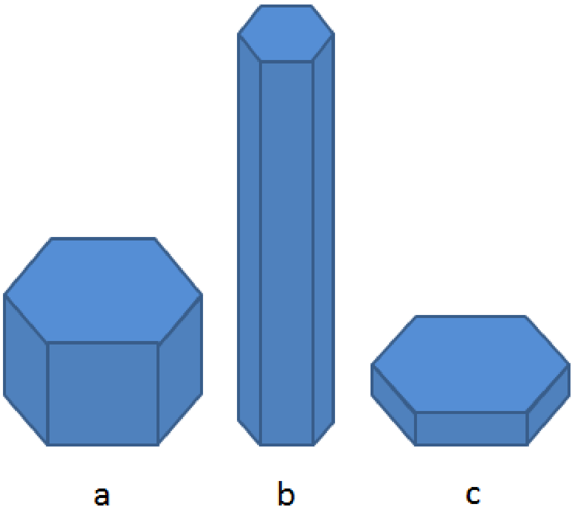


Fig. 40. Types of Crystal Habits. a – Cubic. b – Column. c – Tabular.

In addition to temperature, such factors as pressure, composition of the solutions from which the minerals crystallize, and variation in mineral composition may affect crystal forms. The presence of impurities in a mineral may also modify the habit of a crystal.

3.3. Physical Properties of Minerals

Each mineral possesses certain physical properties or characteristics by which it may be recognized or identified. These properties include color, streak, hardness, luster, specific gravity, cleavage and fracture, play of colors, transparency, magnetism, etc.

Color is one of the first things that is noticed about a mineral. But the same mineral may vary in color from one specimen to another. Only some minerals have relatively constant colors. Azurite is always blue, malachite – green, cinnabar – red, pyrite – yellow. Quartz or tourmaline occur in a wide variety of colors. Very often color variations are due to minor chemical impurities within the mineral. Apatite may be green, light-blue, grey, brown, violate, and white. Quartz's crystals may be colorless (transparent), milky white, black (morione), and smoke-colored (rauch-topaz). Color may be of little use in identifying some minerals. Anyway, the specimen should be examined in natural light. The surface should be fresh, not weathered and dry, not wet. Certain of the metallic minerals will tarnish and the true color will not be revealed except on a fresh surface.

Streak. When a mineral is rubbed across a piece of unglazed tile, it may leave a line similar to a pencil or crayon mark. This line is composed of the powdered minerals. The color of this powdered material is known as streak of the mineral, and the unglazed tile used in such a test is called a streak plate. The streak in some minerals will not be the same as the color of the specimen. A piece of black hematite will leave a reddish brown streak, while a specimen of a pyrite will leave a streak of greenish gray instead of a golden yellow. An extremely hard mineral such as topaz or corundum will leave no streak as it will not be powdered by a softer surface of a streak plate.

Hardness is a very important identifying feature of a mineral. The hardness of a mineral is determined by what materials it will scratch, and what materials will scratch it. The hardness or scratch test may be done with simple testing materials carried in the field. For greater accuracy one may use the scale of hardness called Mohs' scale. In establishing this scale, Mohs, the German mineralogist, assigned each of the reference minerals a number. He designated talc, the softest in the series, as having a hardness of 1. The hardest mineral, diamond, was assigned a hardness of 10. Mohs' scale, composed of the tem reference minerals arranged in order of increasing hardness, is as follows:

No 1 – Talc (softest)
No 2 – Gypsum
No 3 – Calcite
No 4 – Fluorite
No 5 – Apatite
No 6 – Feldspar
No 7 – Quartz
No 8 – Topaz
No 9 – Corundum
No 10 – Diamond (hardest).

Note that Mohs' scale is so arranged that each mineral will be scratched by those having higher numbers, and will scratch those having lower numbers.

It is also possible to test for hardness by using the following common objects (Table 8):

Item	Hardness
Fingernail	About 2 ½
Copper penny	About 3
Glass	5 - 5 ½
Knife blade	5 ½ - 6
Steal file	6 ½ - 7

Each of the above items will scratch a mineral of the indicated hardness: the fingernail will scratch talc (hardness of 1) and gypsum (hardness of 2), but would not scratch calcite which has a hardness of 3. Minerals with hardness of 6 and more will leave scratches on a steel blade of a knife and on a glass.

Luster. The appearance of the surface of a mineral as seen in reflected light is called luster. Some minerals shine like metals, for example silver or gold. These are said to have metallic luster. Other lusters are called nonmetallic. The more important nonmetallic lusters and some common examples are given below:

Adamantine – brilliant glossy luster: typical of diamond
Vitreous – glassy, looks like glass: quartz or topaz
Resinous – the luster of resin: sphalerite
Greasy – like an oily surface: nepheline
Pearly – like mother-of-pearl: talc
Silky – the luster of silk or rayon: asbestos or satin-spar gypsum

Dull – as the name implies: chalk or clay.

Submetallic luster is intermediate between metallic and nonmetallic luster. The mineral wolframite displays typical submetallic luster. Here too, one must take into consideration such factors as tarnish, type of lighting, and general condition of the specimen being examined.

Specific Gravity is determined by comparing the weight of the mineral specimen with the weight of an equal volume of fresh water. A specimen of galena (specific gravity about 7.5) would be about 7.5 times as heavy as the same volume of water. Specific gravity of most minerals changes from 0.9 to 2.3 g*cm^{-3}.

Minerals with specific gravity up to 2.5 (halite, gypsum, sulphur, etc.) are called light minerals, those with specific gravity within 2.5-4.0 (quartz, calcite, orthoclase, etc.) are middle-weight minerals, and those with specific gravity over 4.0 are heavy ones (like hematite, magnetite, cinnabar and pyrite).

Cleavage and Fracture. Mineral crystals will break if they are strained beyond their plastic and elastic limits. If the crystal breaks irregularly if is said to exhibit fracture, but if it should break along surfaces related to the crystal structure. It is said to show cleavage. Each break or cleavage plane is closely related to the atomic structure of the mineral and designates planes of weakness within the crystal. The number of cleavage planes present and the angles between them are constant for any given mineral. Cleavage is, therefore, useful in mineral identification. The mineral galena cleaves in three planes (directions) at right angles to one another. Calcite has three cleavage planes that are not at right angles to one another. Hence, galena is said to have cubic cleavage, and calcite rhombohedral cleavage. Four varieties of cleavage are recognized: perfect (minerals easily split into parts with smooth surface): micas, some specimen of gypsum and graphite; nearly perfect (minerals split or break into smooth-faced fragments, but not so easily): orthoclase, halite, calcite, etc; moderate (angite and amphiboles), and nonperfect (not easily detectable): apatite, beryl, etc. Many minerals do not possess any cleavage, but their broken surfaces (fractures) may be of value in identifying them. There are several types of fractures; some of the more common types are:

Conchoidal – the broken surface of the specimen shows a fracture resembling the smooth curved surface of a shell. This type of fracture is typical of chipped glass, quartz, and obsidian.

Splintery or Fibrous – fibers or splinters are revealed along the fracture surface: pectolite.

Hackly – fracture surfaced marked by rough jagged edges: copper, silver, and certain other minerals.

Uneven – rough irregular fracture of surface. This type is common in jasper, a variety of quartz and many other minerals.

Even – as the name implies: magnesite.

Earthy – as the name implies: kaolinite.

Other physical properties may also aid greatly in identification. Play of colors: variations in color when viewed from different angles: labradorite. Asterism: the mineral exhibits a star-like effects when viewed either by reflected or transmitted light: certain specimen of phlogopite or the star-like – sapphire. Transparency or lack of it: opaque – no light passes through the mineral: galena, pyrite, and magnetite. Transluscent – light passes through the mineral but an object cannot be seen through it: chalcedony and certain other varieties of quartz. Transparent – light passes through the mineral and the outline of objects can be clearly seen through it: halite, calcite, clear crystalline quartz.

Magnetism: a mineral is attracted to an iron magnet: magnetite, or lodestone, and pyrrhotite. Some minerals may glow or emit light by exposure to ultraviolet rays. Exposure to X-rays, cathode rays, or radiation from radioactive substances can cause luminescence too. A mineral exhibiting phosphorescence will continue to glow after the cause of excitation has been removed. Rhombohedral calcite crystal exhibits double refraction.

The mineral may occur in nature in various forms. Ideal monocrystals are rare (quartz, rhombohedral calcite, micas, pyrite). Doubles of crystals are also rare: gypsum, quartz, orthoclase. More often occurring are concretions, geodes, crystal and earthy masses.

3.4. Origin and Classification of Minerals

The majority of rock-forming minerals are silicates (feldspars, pyroxenes, amphiboles, micas, clay, minerals, etc.) and other compounds of silicon (quartz, opal, etc.). The minerals are formed in nature under a variety of conditions. The entire set of the processes of mineral formation has been divided into three groups depending on the source of energy utilized in these processes: endogenic, exogenic, and metamorphogenic.

Endogenic processes of mineral formation are the processes of internal dynamics. The minerals of igneous rocks were formed by these processes. Igneous rocks have solidified from an original molten state, the Latin word ignis meaning “fire”. The temperature within the earth are exceedingly hot and many rocks and minerals exist in a molten condition called magma (from Greek – dough). These magmas are large bodies of molten rock deeply buried within the earth. Magmatic formation of minerals occurs at high temperatures (700-1600°C) and pressure (dozens and hundreds of MPa). Rising up within the crust the magma gradually gets cooler and solidifies forming the crystals of various minerals (crystallizing differentiation). At the beginning of this process (at 1600°C) some hardmelting minerals crystallize, such as magnetite (Fe_3O_4), chromite ($FeO*Cr_2O_3$) and apatite [$Ca_5(PO_4)_3(Cl,F,OH)$]. Then iron-magnesium silicates appear: olivine, augite, hornblende, biotite, etc. Further on, at 800-900°C, muscovite and basic plagioclases form their crystals. Then, if some free silica still remains in solidifying magma, there takes place the crystallization of quartz.

At the next stage the pegmatitic processes of mineral formation occur and pegmatitic dikes and veins commonly result from the residual magma being injected into cracks and joints in the rocks of already crystallized magma. The pegmatitic processes occur at 500-600°C. The pegmatites contain large crystals of feldspars overgrown with smaller crystals of quartz and other minerals. Pegmatitic veins very often contain some large crystals of microcline, apatite, topaz, beryl, tantal, tourmaline and the minerals of some rare elements. Such veins are the only industrial source of a muscovite (white mica).

During further solidification of magma the amount gases in it becomes excessive and they begin to evolve interacting with each other and with newly formed minerals. Thus begins the pneumatolitic processes of mineral formation. Many chemical elements leave magma in this way. The gases that are emitted from the molten magma are composed largely of water vapor, with varying amounts of carbon dioxide, hydrogen sulfide, and chlorine. Pneumatolitic rocks are composed of quartz, muscovite, topaz, rare elements; some valuable ores are formed in this way, containing tungsten, tin and molubdenum.

The further cooling of magma at high pressure and temperature up to 350-370°C is accompanied by the flow of overheated (but liquid) water which, possessing expressed acidic properties, is capable to dissolve many substances. Such (hydrothermal) processes give birth to the formation of noble and colored metals (in native states or composing certain polymetallic ores).

Sometimes magmatic materials are poured out upon the surface of the earth, as, for example, when lava flows from a volcano. The magma loses its gases and undergoes relatively rapid cooling. A part of its gaseous components (HF, HCl, H_2S, SO_2, P, etc.) evaporate. At low temperatures, such components can turn from the gaseous to solid state. In this way the native sulfur as well as boron and some other minerals may form in the volcanic craters (exhalationic process of mineral formation). Endogenic minerals are, thus primary minerals which form at high temperatures and are derived from igneous rocks.

Exogenic processes of mineral formation are the processes of external dynamics. They take place at or near the surface of the earth. They are subjected to the influence of solar energy, temperature, oxygen and other components of the atmosphere, water and aqueous solutions, living and dead organisms. They occur at relatively moderate temperatures and atmospheric pressure.

Weathering, the process whereby rocks undergo natural chemical and physical change at or near the surface of the earth is one of the most important of all geologic processes. It provides much of the material from which sedimentary rocks are formed, is important in the shaping of surface land forms, and is responsible for the formation of soil. Rock fragments produced by weathering are removed by erosion – the loosening and carrying away of rock debris by natural agents. Weathering and erosion are constantly at work, wearing away the rocks of the earth's surface.

Physical, or mechanical, weathering takes place when a rock is reduced to smaller fragments without undergoing a change in chemical composition. This type of weathering, known also as disintegration, may be the result of a variety of physical forces. Frost action, alternate heating and cooling, organic activities – all are the most potent agents of physical weathering.

When water freezes in the cracks or pores of rocks it expands exerting sufficient pressure to break down the rock. In frost wedging these pressures are directed laterally; in frost heaving, which usually occurs in unconsolidated rocks, they are exerted upward.

In some areas of the earth, particularly in certain mountainous regions, the rocks are subjected to drastic temperature changes almost daily. Rocks of a high mountain peaks expand as they are heated in the daytime, and contract when subjected to freezing temperatures at night. This process, repeated over long periods of time, will cause small cracks and crevices which permit other agents of weathering, such as frost wedging or solution (see below) to attack the rock. In rocks with well formed distinct crystals of various minerals possessing varying coefficients of volume and linear expansion, the temperature variations cause the disintegration of a rock into the grains of the minerals composing it.

The activities of various organisms also promote rock disintegration and chemical transformation of primary minerals into secondary ones. This is organic or biological weathering. Some geologists are still undecided as to indentifying it among the types of weathering, as its mechanisms, in their essence, belong to either physical or chemical weathering. Anyway, biological weathering is very potent and soil formation would be impossible without it. Tree roots, which frequently grow in rock crevices, can exert sufficient pressure to force rock fragments apart. Some bacteria (Bac. Extorquens) are capable to disintegrate and decompose silicates. Lichens, exuding organic acids, can decompose (dissolve) feldspars. The fungi are no less efficient decomposers of alumosilicases and iron and aluminum oxides. Some burrowing animals as rodents, worms and ants, bring to the surface rock particles to be exposed to the action of weathering. All these organic activities, taking place in the masses of disintegrated rocks, signify the beginning of the primary process of soil formation. The roots and residues of higher plants accelerate such processes very much.

Chemical Weathering, or decomposition, produces a chemical breakdown of the rock, which may destroy the original minerals and produce new ones. Physical weathering simply produces smaller fragments of the parent rock; chemical weathering produces rock materials that are basically different form the original rock. Although chemical changes occur in a variety of ways, the more common processes of decomposition are oxidation, hydration, carbonation, solution, etc.

Oxidation occurs as oxygen, assisted by moist air, combines with minerals to form oxides. Rocks and minerals containing iron compounds are especially susceptible to this type of decomposition. The oxidation of iron compounds, which produces rust, is also responsible for the color of many red, yellow, and brown rocks and soils. Certain iron compounds (for example, pyrite) form acids when oxidized:

$$2FeS_2 + 7O_2 + H_2O = 2FeSO_4 + H_2SO_4$$

$$12FeSO_4 + 6H_2O + 3O_2 = 4Fe_2(SO_4)_3 + Fe(OH)_3$$

$$2Fe_2(SO_4)_3 + 9H_2O = 2Fe_2O_3{*}3H_2O + 6H_2SO_4$$

The acid attacks the rocks and thereby hastens the process of decomposition. The reaction of a feldspar with water may illustrate the hydrolysis:

$$KAlSi_3O_8 + H_2O = KOH + HAlSi_3O_8$$

An acid environment stimulates the reaction of water with minerals and is one of the most important weathering reactions:

$$\underset{\text{albite}}{4NaAlSi_3O_8} + 18H_2O + 4H^+ = \underset{\text{kaolinite}}{Al_4(OH)_8[Si_4O_{10}]} + 8H_4SiO_4 + 4Na^+$$

In the reaction a primary mineral, albite, is converted to kaolinite. The first-formed kaolinite particles are very small. They increase in size as additional ions attach or join the crystal, resulting in a gradual increase in particle size. Clay particles like kaolinite are called clay minerals. Some clay minerals tend to be resistant to further weathering. A process described above is sometimes called "kaolinization". Chemical weathering of the most abundant primary minerals (orthoclase, albite, etc.) gives birth to a number of secondary minerals: salts, oxides (hydroxides) and clay minerals. The kaolinite, even though very resistant to further weathering, may be decomposed and disappear from the rock or soil:

$$Al_4(OH)_8[Si_4O_{10}] + 10H_2O = \underset{\text{gibbsite}}{4Al(OH)_3} + 4H_4SiO_4$$

Hydration of minerals (their chemical combination with water) produces hydrous silicates and hydrous oxides. Some examples of this are the conversion of anhydrite to gypsum and the reaction between hematite and water to produce limonite:

$$\underset{\text{anhydrite}}{CaSO_4} + 2H_2O = \underset{\text{gypsum}}{CaSO_4{*}2H_2O}$$

$$2Fe_2O_3 + 3H_2O = \underset{\text{limonite}}{2Fe_2O_3{*}3H_2O}$$

In addition to the chemical effect of hydration, there is a physical expansion of the minerals during this process. Thus the hydration of anhydrite increases its volume by 50-60%. Besides, its solubility in water at 20°C increases from practically zero to 2.6 $g{*}l^{-1}$.

Carbon dioxide (CO_2), which is generally present in air, water, and soils, commonly unites chemically with certain rock minerals, greatly altering their composition. Substances produced in this manner (carbonates and bicarbonates) are relatively soluble and therefore easily removed and carried away. In addition, the union of carbon dioxide and water produces carbonic acid (H_2CO_3), an effective agent in attacking such minerals as calcite and dolomite:

$$CaCO_3 + H_2O + CO_2 = Ca(HCO_3)_2$$

The joint action of oxidation, hydration and carbonation is a powerful agent of weathering of amphiboles and pyroxenes:

$$\underset{\text{augite}}{4CaFeSi_2O_6} + O_2 + 4H_2CO_3 + 6H_2O = 4CaCO_3 + \underset{\text{goethite}}{4FeO(OH)} + 8H_2SiO_3$$

Reduction also plays a significant role in the weathering of minerals containing reducing elements in an oxidized form. But this process involves the participation of chemotrophic microorganisms and proceeds in the deficit of oxygen.

The upper part of the earth's crust where the processes of weathering are noticeably active is sometimes termed as the "cortex of weathering". Soil scientists consider the soils to be new formations within the cortex of weathering. They prefer to classify the cortex of weathering by dividing it into (1) sialitic and (2) alitic.

Sialitic cortex of weathering is widespread in the regions with moderately humid climate and not so high temperatures. Clay minerals which form as a result of weathering, mainly smectites and hydromicas, are resistant to any further changes and tend to accumulate in this cortex:

$$2K[AlSi_3O_8] + 4H_2O + 2H^+ = \underset{\text{pyrophylite}}{Al_2(OH)_2[Si_4O_{10}]} + 2K^+ + 2H_4SiO_4$$

Primary minerals resistant to weathering also accumulate in parent materials and soils. Quartz becomes the most abundant mineral in sedimentary rocks, parent materials and mineral soils.

Alitic cortex of weathering develops in humid and relatively hot climates of tropics and subtropics. Clay minerals and hydromicas cease to be resistant to weathering under such conditions and turn into hydroxides and oxides of aluminum and iron. Even kaolinite decomposes, resulting in the formation of gibbsite and silicic acid. Gibbsite is a mineral more resistant to weathering then the kaolinite. The loss of silicic acid by leaching is resulting in the progressive loss of silicon. There is a progressive increase in the accumulative of aluminum, which tends to be incorporated into resistant secondary minerals that accumulate in the soil.

Metamorphogenic processes of mineral formation. Metamorphic rocks are rocks (originally either igneous or sedimentary) that have been buried deep within the earth and subjected to high temperatures and pressures. During the process of metamorphism the original rock undergoes physical and chemical alterations which may greatly modify its texture, mineral composition, and chemical composition.

Minerals are formed with dense crystal structures. In the deeper zones of metamorphism the minerals undergo dehydration, deoxydation and decarbonation. Thus opal may be turned into a chalcedone, and chalcedone, in its turn, into quartz; limonite changes into a hematite, while hematite, losing a part of its oxygen, turns into a magnetite. Carbonates change to silicates losing their carbonic acid. Many minerals of metamorphic origin (micas, talc, chlorite, hornblende, etc.) acquire a characteristic platy, leafy and occasionally needle-shaped structure. Limestone may be metamorphosed into marble, and sandstone into quartzite. Coal changes to graphite. Magmatic fluids commonly introduce new elements and compounds which will modify the chemical composition of the original rock and result in the formation of new minerals.

Classification of minerals. Although different classification schemes could be used, mineralogists have found that minerals with similar chemical compositions share a great number of other properties than minerals grouped by other criteria. Thus, minerals are first divided into classes depending upon the dominant anion or anionic group. The main classes are:

1. native elements;
2. oxides and hydroxides;
3. carbonates;
4. phosphates;
5. sulfates;
6. nitrates;
7. halides;
8. sulfides, and
9. silicates.

These classes are subdivided into subclasses on the basis of chemical and structural similarities. Commonly occurring mineral classes, particularly the silicates, are covered here in more detail than the others.

3.5. Native Elements

The minerals of native elements, composed of only one element, are not very abundant in the earth’s crust: some 30 minerals occur in it more or less frequently. First of all, these are noble metals (Au, Ag, Pt, etc.), some colored metals (Cu, Hg, Bi, Pb), nonmetals (As, C, S) and gaseous elements (O, N, He, etc.). In Ukrainian mineralogy, unlike in that of other countries, the gaseous and liquid minerals are considered the minerals too. And there is a proper reason for it. Some minerals occur as allotropic varieties (for example, carbon in the forms of diamond and graphite).

But the native elements are very important for the industries and other spheres of human activity. The most widespread of them are graphite and sulfur.

Graphite-C. The mineral’s name is derived from the Greek “Υραφω” – to write. The mineral is of grey, sometimes black color, black streak, metallic to dull luster. It seldom occurs in the form of distinct crystalline formations (hexagonal tablets). Hardness 1: specific gravity 2.2; greasy to touch; perfect cleavage; a good conductor of electricity. The mineral is of igneous (pegmatitic) or metamorphic origin. As a rock-forming mineral it occurs in metamorphic slates of a very fine texture in which mineral crystals cannot be detected with the naked eye; it splits readily into thin even slabs. It also occurs in gneiss, a rock characterized by alternating bands of darker minerals – graphite among them.

Igneous deposits of a graphite are represented by massive accumulations, in which graphite is associated with feldspars (microcline or nepheline). In Ukraine, the largest deposits were formed by metamorphosis in Volyn, Kiev Region, Left Bank of the Bug, Krivyi Rig and others.

Graphite is commonly used for the manufacture of crucibles for the smelting of rare metals and some special grades of steel. Electrodes and contacts of electric machines are made from it, as well as some lubricants and paints. Lead pencils do not contain lead, but a graphite.

Sulfur-S is the only mineral of the class that possesses a molecular structure. Very often it occurs in a chemically pure state, but sometimes it contains the impurities of As, Se, Te and may be polluted by bitumens, clay and carbonates. It is of straw-yellow color when pure but may also be honey-yellow or yellowish-brown, reddish, greenish, grey, brown and even black; streak-colorless or pale-yellowish; luster resinous to greasy, but may be adamantine on the faces of crystals; hardness: 1 ½ to 2 ½; specific gravity: 2.1; fracture: Conchoidal to uneven.

This mineral is found in volcanic rocks, around hot springs, and associated with salt domes. It can be inflamed from a match and to burn with a blue flame and the emanation of SO_2. Sulfur may be of volcanic and exogenic (sedimentary) origin. Pneumatolitic processes like: $SO_2 + 2H_2S \rightarrow 3S + 2H_2O$ occurs in the crevices of igneous rocks. The industrial deposits of sulfur are connected with metallic (sulfide) ore and carbonates, silicates and sulfates of calcium. Much of the United States production is from the "cap rock" of certain salt domes in the Gulf Coast region of Texas and Louisiana.

Sulfur is used as a beaching agent, in making paper, sulfuric acid, gunpowder, matches, insecticides, and medicines. It is also used in the process of vulcanizing rubber.

3.6. Oxides and Hydroxides

The minerals formed of a union of various elements with oxygen belong to the class of oxides. If water also enters their composition, they are called hydroxides. Quartz is the most abundant of these minerals (12.5% by weight of the earth's crust). In nature silica (SiO_2) occurs as seven distinct polymorphs: quartz, cristobalite, tridymite, coesite, stishovite, lechatelierite (silica glass), and opal; the latter two are amorphous. Of these minerals, quartz is most abundant in soil environments, while disordered cristobalite commonly occurs in soils. Opal is a hydrated “amorphous” silica that comprises the bulk of diatomaceous rocks of biogenic origin.

Quartz is an important part of many igneous rocks and is common in sedimentary and metamorphic rocks. Quartz may occur in combination with other minerals, or, as in the case of pure sandstones and quartzites, may be the only one present.

Pure quartz is composed of silicon dioxide (SiO_2), but certain varieties contain impurities such as iron or manganese. These impurities are responsible for the varied colors of certain types of quartz. Quartz occurs in crystalline aggregates or in irregular grains or masses. The term cryptocrystalline (literally “hidden crystals”) is used to describe quartz varieties in which crystals are not evident.

Crystallizing in the hexagonal system, quartz commonly forms six-sided crystals with pyramidal ends. It may be colorless, white, rose, violet, smoky, gray, or a wide variety of other colors. It has a vitreous to greasy luster, a hardness of 7, specific gravity of 2.65, and conchoidal to uneven fracture. Quartz may be transparent to opaque and leaves a white or very pale colored streak. Some of the more common crystalline varieties of quartz are amethyst, milky quartz, smoky quartz, pink quartz, rock crystal quartz (see Fig. 41).


a b

Fig. 41. Well-developed Hexagonal Quartz: a – Crystals with Pyramidal Ends; b – Rock Crystal Quartz.

Massive or cryptocrystalline varieties include agate, chalcedony, chert, flint and jasper. Quartz is apt to occur almost anywhere, and most sands are composed largely of quartz fragments. Quartz is used in the manufacture of electronic equipment and some varieties – in making lenses and prisms. Other types of quartz are valuable as semiprecious stones or gems. Sandstone is used as building stone, and quartz sands are used as abrasives and in making concrete and glass.

The mineral is found in essentially all soils and parent materials. It often constitutes the major portion of all sand and silt fractions and is a major component of the coarse clay fraction of many soils.

Opal is a naturally occurring hydrous (SiO_2*nH_{20}), x-ray amorphous silica polymorph. It occurs in most soils only in minor quantities. It may be of either organic or inorganic origin. Organic bioliths and inorganic siliceous accumulations may be common in duripans. Biogenic opal, which includes bodies with characteristic shapes, sizes, and forms that are deposited in or secreted by plants and animals, may be found in soils and geologic sediments as residue products.

Iron oxides and hydroxides are among the most valuable products known to man as are all the other metallic or ore minerals. Iron, man's most useful and important metal, is obtained from a variety of iron minerals. Some of them tend to accumulate in the cortex of weathering, parent materials and soils. Anhydrous iron oxides include magnetite (Fe_3O_4) and hematite (Fe_2O_3). The most widespread among the hydroxides of iron is limonite ($2Fe_2O_3*3H_{20}$).

Magnetite (Fe_3O_4) is a combination of ferric (Fe_2O_3) and ferrous (FeO) oxides. It occurs among the metamorphic rocks (gneiss or slates), where it is formed from hematite. It is strongly attracted by a magnet. Magnetite contains 75% iron. Its color is mostly black (iron-black) and so is its streak; hardness: 5 ½ to 6 ½; luster: metallic to submetallic; fracture: conchoidal to uneven; specific gravity: 5.0 to 5.2. A variety of magnetite that acts as a magnet is known as lodestone. It occurs in nature as crystals of a cubic system (octahedral). Though igneous and metamorphic origin predominates, it can also form as are result of weathering of angite, feldspar, and chlorite.

Hematite (Fe_2O_3) is the most important ore of iron. One of the world's most common minerals, it occurs in massive black beds and in scaly schistose rocks. Hematite is mostly sedimentary in origin, and most deposits have been altered and enriched by subsequent solutions. Sometimes it may form by the metamorphosis of a limonite (dehydration):

$$2Fe_2O_3*3H_2O \rightarrow 2Fe_2O_3\ (-3H_2O)$$

or by the weathering (oxidation) of a magnetite:

$$2Fe_3O_4 + \frac{1}{2} O_2 = 3Fe_2O_3$$

Occasionally hematite forms what is known as iron rose: the scales of a mineral combine very much like the petals of a rose. Oolitic (bean) formations are also encountered in the form of earthy masses. Physical properties: color: gray, reddish brown to iron black; hardness: 5 to 6 ½; specific gravity: 4.9 to 5.3; luster: metallic to earthy; streak: red (regardless of color), hence the name from Greek *γεματίκοσ* – “bloody”; fracture: splintery to uneven. Large deposits of hematite are found near Krivoy Rog (Ukraine). In the United States large deposits of this ore are found in Minnesota, Alabama, and Michigan.

Limonite ($2Fe_2O_3*3H_2O$) is not so much a mineral as it is a rock. A term is used to refer to several comingled hydrous iron oxides. It occurs in compact or earthy masses and is a relatively common iron ore. The mineral is of exogenic origin and, therefore, belongs to the secondary minerals. It is an amorphous mixture of colloidal particles of iron oxide and adsorbed water, so its chemical composition is not definite. Physical properties: color: yellow, brown, or black; hardness: 1 to 5 ½; specific gravity; 3.4 to 4.0; luster: dull earthy; streak: yellow-brown; fracture: uneven. Though the crystal form is not typical to limonite, it sometimes forms pseudomorphoses like small cubes on a pyrite. Among the forms occurring in nature there are “kidneys”, concretions, geodes, and oolites (“beans”). As an ore, limonite is used in the manufacture of pig iron and steel. Some clayey limonites and ironous clays are used for the manufacture of paints: yellow and red ochres.

Iron oxides and hydroxides in soil environments. Iron oxides occurs, at least in small amounts, in nearly all soils. The absence of Fe oxides, or their occurrence in only trace amount, usually is due to strong reduction and removal from the profile or the absence of Fe-bearing primary minerals in soil parent materials.

Magnetite occurs from trace to moderate quantities in soils and sometimes makes up a significant portion of the “heavy” mineral fraction. This is especially true for young soils influenced by basic igneous rocks – e.g. basalt.

Goethite (FeOOH) generally is considered the most widespread Fe mineral in soils, occurring under wide-ranging conditions. It is a significant component of most Oxisols (“red earth”). Goethite is the most common Fe oxides identified in soil concretions and modules, which often form in soils that are wet at least part of the year. Goethite, and other Fe oxides containing Fe^{3+}, are unstable in a reducing environment. The solubilized Fe^{2+} may be translocated to another part of the profile, where there is a more oxidizing environment to form mottles or concretions.

Hematite occurs in a great variety of soils, but more of it is present in soils of tropical and warm temperate regions than in those of colder areas. Hematite is an effective soil – pigmenting agent. It has been correlated with red color development. Hematite often is a constituent of concretions and nodules. Limonite, or rather ferrihydrite, is associated with bog iron deposits and occurs in spodic (illuvial) horizons of sand-textured soils (like Ukrainian sod podzolic soils). In such soils, iron and aluminum oxides (sesquioxides) are released in weathering and complex or chelate with organic matter and are transported to the subsoil, resulting in the formation of an illuvial horizon.

Aluminum oxides and hydroxides. Aluminum, one of the more important metals of industry, is derived primarily from bauxite, a hydrous aluminum oxide ($Al_2O_3*2H_2O$). Corundum (Al_2O_3) is also a widespread mineral of blue or yellow colors and barrel-shaped crystals. Its hardness is 9 (scratches all the minerals except almond); specific gravity: 4; luster: bright vitreous. The mineral contains 53 % Al and some impurities like Cr, U, Fe and Ti. In modern technology, corundum is used as abrasive material and in cutting instrument. As a rule, this mineral occurs in a pure state, but may also be encountered as a rock-forming mineral. Its origin is connected with the cooling magmas and the metamorphism. It often occurs in elluvial deposits occur in the Urals (Ilmen and Cherry Mountains). Corundum is also a fire-resistant material. Its transparent varieties are employed in precision instrument manufacture, optics and jewelry (ruby).

Bauxite ($Al_2O_3*2H_2O$) is actually a mixture of minerals. It occurs in earthy, claylike masses, or in a pisolithic form as rounded concretions in a claylike matrix. Physical properties are: color: white, yellow, brown, red; hardness: 1 to 3; specific gravity: 2.0 to 2.5; streak: like the color. Diaspore ($Al_2O_3 * H_2O$) and hydrargillite ($Al_2O_3*3H_2O$) are the main minerals composing bauxite, which is of secondary origin connected with microbiological weathering of feldspars and the other aluminosilicates. Bauxite origin is also connected with alitic cortex of weathering in humid tropics and subtropics. It is the main aluminum ore, the closest to Ukraine deposits of which are on the territory of a Russian Federation (Tikhvin and in the Urals, along the Chusova river).

Corundum, formed mostly in metamorphic rocks, is occasionally identified in the heavy-mineral fraction of soils, as it was listed as one of the more resistant minerals in the weathering sequence.

Gibbsite [$Al(OH)_3$] is by far the most common Al-hydroxide in soils; sometimes it is the predominant mineral in the clay fraction. The usual pathway proposed for the gibbsite formation is by the desilication of kaolinite. Gibbsite has been detected in intensively weathered soils of Australia; its occurrence in more arid environments was considered to be a relict of more severe weathering, when the climate was more humid.

Boehmite (AlOOH) is much rarer in soils than gibbsite. It has been more commonly identified in intensively weathered soils.

3.7. Carbonates

Calcium and magnesium carbonates are very common in the earth's crust. They compose many sedimentary and metamorphic rocks: limestones, chalk, marble, dolomite and marl. Calcite and magnesite are important components of loesses and loesslike loams – the most widespread parent materials of soils in Ukraine.

Calcite ($CaCO_3$) is composed of calcium carbonate and is the most common member of the calcite group. It occurs in many sedimentary and metamorphic rocks, and is the primary constituent of most limestones. Calcite occurs in crystalline, granular, or chalky masses, as a vein mineral, in cave and spring deposits, and in the shells of certain animals (corals, snails, clams, and others).

Perfect rhombohedral crystals, forming in the hexagonal system, are common. Calcite is typically colorless, white or yellowish, but the presence of impurities may bring about a wide variety of colors. Other properties: hardness: 3; specific gravity: 2.72; luster: vitreous to dull; streak: white to grayish; cleavage: perfect rhombohedral in three directions at oblique angles. Calcite will effervesce or "fizz" in cold dilute acids; some forms are fluorescent; and certain clear calcite crystals have the property of double refraction. An object when viewed through such a crystal will appear double (see Fig. 42).

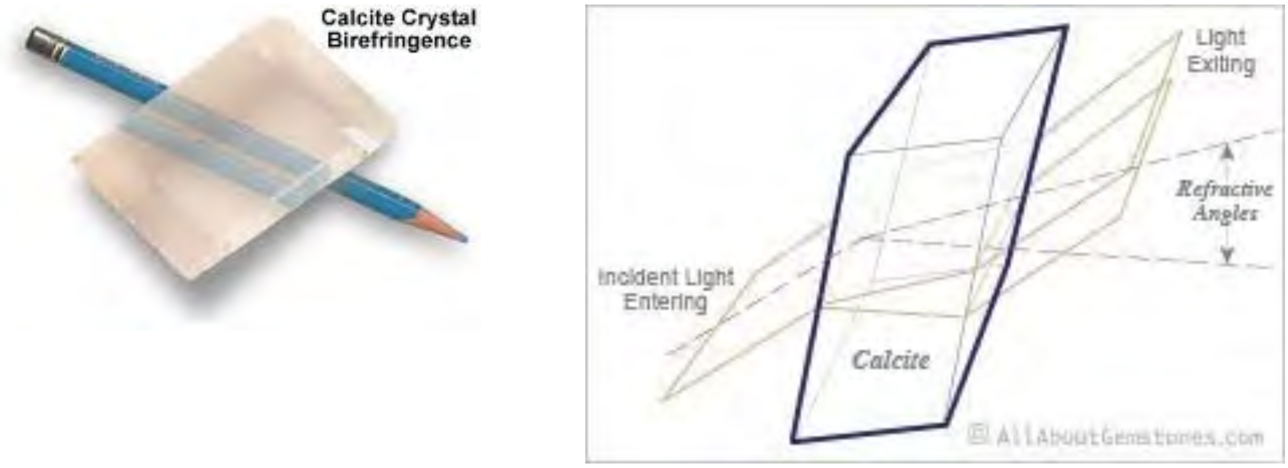


Fig. 42. Rhombohedral calcite crystal exhibiting double refraction.

Some of the more common varieties of calcite are Iceland spar, dogtooth spar, chalk, travertine (including the calcacerons tufa deposited by springs, stalactites and stalagmites formed in caves), and limestone.

Magnesite ($MgCO_3$) is much less widespread than calcite. It may be white, gray or yellow in color, its streak remaining white. Hardness: 4, specific gravity: 3, luster vitreous, dull and occasionally silky. Occurs in nature mostly as coarse-grained marble-like or chalk-like amorphous masses. Effervesces only in hot HCl; is used for the production of special bricks.

Dolomite [$CaMg(CO_3)_2$] is common in sedimentary rocks, where it is often mixed with calcite. It may also occur in association with many ore deposits and in veins or cavities of some igneous rocks. Dolomite differ from calcite in that it is slightly harder (3 ½), is only mildly affected by cold dilute acids, and may include crystals with curved faces. Its major uses are as the building stone, in the manufacture of cement, and as a source of magnesium. As an agronomical liming material (aglime), ground dolomite is used for the neutralization of excessive soil acidity.

Calcium and magnesium in parent materials and soils belong to the macroelements of plant nutrition. Aragonite, like calcite, is composed of calcium carbonate, but it differs from calcite in that it is less stable and crystallizes in the orthorhombic system.

Aragonite has a higher specific gravity (2.9) and is somewhat harder (3 ½) than calcite. It occurs as a secondary mineral in cavities of limestone, as a deposit around hot springs and geysers, in cave deposits, and in the shells of certain animals, such as clams or corals. Aragonite is used for the same purposes as calcite.

Carbonates in soils environments. Calcite is extremely common soil mineral in dry (subhumid to arid) regions. It may be inherited or authentically formed. It often makes up much of the soil mass, especially in thick calcic horizons, some of which are made up of more than 900 g kg^{-1} calcite.

Carbonates also are common constituents of soils in humid regions. They occur as: (1) inherited components in profiles developed in calcareous materials; (2) anthigenic precipitates at the capillary fringe where there is a shallow water table and underlying calcareous sediments; and (3) carbonates precipitated from percolating water in slowly permeable materials or at the contact between materials of highly contrasting permeability.

3.8. Sulfates

Gypsum ($CaSO_4*2H_2O$) is a very common mineral. A product of evaporation, it occurs in thick-bedded deposits in many parts of the world. Sometimes the origin of this mineral may be hypergenic (hydration of anhydrite deposits).

Gypsum is colorless or white (but may be a variety of colors if impurities are present), has a hardness of 2, a specific gravity of 2.3, is vitreous to pearly or silky in luster, has a white streak, and is transparent to opaque. It crystallizes in the monoclinic system and has one perfect micaceous cleavage, yielding thin leaves or plates. There are two other less perfect cleaves. Among the more common varieties of gypsum are selenite, satin spar, alabaster and rock gypsum. The aggregates of crystals may be grainy, massive in the form of roses, fibreous masses and coneretions. Gypsiferous horizons may be observed in the loesses of the South of Ukraine at the depth of 1.5-2.4 m (southern Chernozems) and 1.0-1.5 m (dark-chestnut and chestnut soils of the Arid Steppe).

A mineral of great economic importance, gypsum is extensively used in the manufacture of plaster board (sheet rock), plaster of Paris, Portland cement, paint, glass, porcelain, and crayons. The alabaster variety of gypsum is used in statuary and as an ornamental stone. Gypsiferous materials are very important for alkaline (sodic) soil amendment. Calcium of gypsum substitutes the exchangeable sodium of such soils, thus improving the unfavorable properties of the latter. The largest deposit of gypsum in Ukraine is a Bakhmuth one.

Anhydrite ($CaSO_4$) is harder and heavier than gypsum. This mineral typically occurs in massive, fine- to medium-grained, colorless, white, or grayish masses, has a hardness of 3 to 3 ½, a specific gravity of 2.9, a vitreous, pearly, or greasy luster, and leaves a white streak. The mineral is a chemical sediment commonly found in the cap rock of certain salt domes. Anhydrite may fluoresce pink, yellow-green, or blue-white upon heating. It is used in the manufacture of cement and fertilizer, and to a lesser degree as an ornamental stone.

Gypsiferous soils are soils that contain sufficient quantities of gypsum to interfere with plant growth. Soils with gypsum of pedogenic origin are found in regions with ustic (with limited but periodically changing to sufficient available water for plants), xeric and aridic moisture regimes. They do not usually occur under wet climates. V. A. Kovda (1954) states that gypsum accumulation occurs in two ways: by the evaporation of mineralized groundwater and by the precipitation within the groundwater itself.

The origin of sulfate ions in the soil solution is in some circumstances due to the presence of sulfur-rich minerals such as pyrite in the parent or underlying materials. By weathering and oxidation, the sulfur in these minerals is transformed into sulfuric acid which in calcareous soils reacts with $CaCO_3$ to form gypsum.

The solubility of gypsum is 2.6 $g*dm^{-3}$ of pure water at 25°C.

3.9. Phosphates

The three most essential elements for promoting plant growth are nitrogen, potassium and phosphorus. Phosphate rock is a valuable source of phosphorus because it contains a large amount of the mineral apatite. Phosphate minerals form under a wide variety of environmental conditions ranging from silicate melts to natural soils to ocean floor. Various processes: first stages of magma solidification, pegmatitic, hydrothermal, and the external sedimentary processes may result in their formation.

Apatite [$Ca_5(PO_4)_3(Cl, F, OH)$] contains 42 % P_2O_5. The primary phosphorus mineral in phosphate rock is fluoroapatite $Ca_{10}F_2(PO_4)_6$. The color is green or bluish green, and occasionally blue, violete, pink, yellow, brown, etc. Luster: vitreous on facets but greasy on the surfaces of cleavage. Sometimes the mineral occurs in the form hexagonal prisms or needles and much less frequently in the form of short columns with vertical striation on the faces of crystals, grainy or earthy masses and concretions. Physical properties: hardness: 5, specific gravity: 3.2. Occurs in igneous rocks. Large deposits of apatite are in the Kola Peninsula (Russia), whose geologic resources were estimated at 2,000 million tons.

Phosphate rock when acted upon with sulfuric acid produces ordinary superphosphate:

$$\underset{\text{phosphate rock}}{Ca_{10}F_2(PO_4)_6} + 7H_2SO_4 = \underset{\substack{\text{monocalcium}\\ \text{phosphate}}}{3Ca(H_2PO_4)_2} + 7CaSO_4 + \underset{\substack{\text{hydrogen}\\ \text{fluoride}}}{2HF}$$

Treatment of phosphate rock with phosphoric acid produces superphosphate with a higher content of phosphorus:

$$Ca_{10}(PO_4)_6F_2 + 14H_3PO_4 = 10Ca(H_2PO_4)_2 + 2HF$$

Phosphorite – $Ca_3(PO_4)_2$ is impurified calcium phosphate of secondary (sedimentary) origin. It may be of yellow, gray or brown color; specific gravity: 2.7-3.2; hardness: may be like that of apatite or less; luster: dull or semigreasy; streak: gray; cleavage: earthy; does not form any crystals (amorphous). In nature this mineral or, rather, rock is encountered as fine-grained masses, tumor-like fragments in the strata of sedimentary rocks (clays, limestones, sandstones, etc.). Phosphorite concretions of the spherical shape and biogenic origin may be encountered on the ocean floor. Like apatites, phosphorites are used for the manufacture of superphosphate and phosphoric acid. Sometimes a well-ground phosphorite rock is directly used as a fertilizer especially on acid soils where they can dissolve slowly yielding available phosphorus for plant nutrition. Rock phosphate is a preferred source of phosphorus by some organic farmers. In Ukraine, large deposits of phosphorite were discovered in Khmelnitsky Region.

The ferrous phosphate vivianite [$Fe_3(PO_4)_2*8H_2O$] has been identified in buried alluvium and in peats under reducing conditions. It may contain up to 28 % P_2O_5. Crystal varieties possess the following physical properties: hardness: 2, specific gravity: 2.7; color: blue, dark blue, gray-blue or bluish-green. Fresh vivianite crystals are transparent and colorless, but it quickly turns blue as Fe^{2+} changes into Fe^{3+} streak is light-colored but it quickly becomes blue; luster: vitreous to metallic, cleavage: perfect, crystal system: monoclinic. Some aggregates and concretions of vivianite are encountered in iron ores, clays, and accompany the bones and other remains of fossil animals and plants in peats. The most widespread accumulations of this mineral are earthy masses of a bright blue color in peats, some of which contain up to 15% P_2O_5 and are directly used as fertilizers and soil amendments or employed for composting. Considerable deposits of vivianite are in Kerch iron ore deposits.

3.10. Nitrates

Chilean ($NaNO_3$) and Indian (KNO_3) saltpetres are the representatives of this mineral class. They are valuable agronomic ores and can be directly employed as nitrogen fertilizers. But they are no longer important, as the source of essentially all industrial nitrogen, including fertilizer nitrogen, results from the fixation of atmospheric N_2.

3.11. Halides

All the minerals of this class belong to the secondary ones as they were formed by evaporation from aqueous solutions mainly on the bottoms of seas and lakes. They seldom if ever occur in a pure state. Their natural aggregates are nearly always polymineral. Their color is very variable and depends upon the impurities and admixtures.

Halite (NaCl), common rock salt, is colorless or white in pure state but is encountered mostly as grey, pink, orange or bluish crystals of an evident cubic system. Hardness: 2-2.5; specific gravity 2.3; luster: vitreous, but greasy on the weathered surfaces; cleavage: perfect in three directions parallel to the facets of a cube. The mineral may form in soils and be a source of soil salinity. Saline soils contain sufficient salt to impair plant growth. The salts are not only chlorides, but also bicarbonates, and sulfates of sodium, potassium, calcium and magnesium. In Ukraine, the industrial deposits of rock salt are Bakhmuth and Slovyansk-Artemivsk (Donbass). It is much used in the chemical industry as an important source of sodium and chlorine. It is also employed in the tanning of leather, food preparation (60-65%), and in certain refrigerants. These are but a few uses of this all-important mineral which has long been so valuable to man.

Sylvite (KCl) contains 62.5% K_2O. Color: white, blue or yellow-red, depending on impurities; hardness: 2; specific gravity: 1.9; occurs in grainy and dense earthy masses; occasionally forms crystals of a cubic system; is encountered in layers or domes in sedimentary rocks, very often accompanying rock salt deposits; was formed as a chemical sediment from the saturated solutions of sea water in earlier geologic epochs. The most important deposits in Ukraine are those near Kalush. Sylvite is used for the manufacture of potassium fertilizers. Potassium mines were opened in Germany in the late nineteenth century. The potassium deposits, which were formed when ancient seas evaporated, are called evaporite deposits. The salts in the ocean water precipitated as the water evaporated and the salts became more concentrated. Later, these salts were buried under various kinds of overburdens and rocks. Sylvinite is a mixture of KCl and $NaCl$. Langbeinite is a mixture of potassium and magnesium sulfates.

Carnallite ($KCl*MgCl_2*6H_2O$) may be of the same variegated colors. It contains 8.7 % MgO and 14.1% K_2O. It is formed in the deposits of halite and Sylvite in the above described way, but it composes the upper domes of these deposits. It is used for the production of metallic magnesium and potassium fertilizers. Hardness: 2.5; taste: bitter; crystals: rhombic system; peculiar property: highly hygroscopic. In Ukraine it is being mined in Kalush.

The most popular method of processing potassium ores is flotation. The ore is ground, suspended in water, and treated with a flotation agent that adheres to the KCl crystals. As air is passed through the suspension, KCl crystals float to the top and are skimmed off. The material is dried and screened to obtain the proper particle size.

Fluorite (CaF_2) owes its name to Latin “fluere” – to flow in connection with its use as a flux in metallurgy. Sometimes it forms large and beautiful crystals and their aggregates. Color: various shades of grey, green, violete, yellow and brown; seldom it may be colorless; luster: vitreous; hardness 4, friable; specific gravity: 3.2; origin:

endogenic-hydrothermal or metasomatosic; occurs, though not often, in igneous rocks and pegmatites. Large deposits were discovered near the Jurassic stratum on the coast of the Arctic Ocean, in the Central Asia, and near the Lake Baikal. Fluorite is used as raw material in the production of fluorine and its compounds. 70% of its output is used in the metallurgy.

3.12. Sulfides

Sulfides of various metals: Fe, Cu, Pb, Zn, Hg, Ag, etc. belong to this class. They are important metallic ores. Their origin is connected with hydrothermal processes. Sometimes the igneous rocks contain them (sulfides of Cu, Ni and Fe) and sometimes they may be the products of weathering and occur in sedimentary rocks. Having arrived on the surface of the earth, these minerals are attacked by the weathering processes. Various secondary products of oxidation, including ochres, may be formed. Sulfide accumulation and the subsequent formation of acid sulfate soils are limited primarily to coastal margins. Pyritic tailings from hard rock mining operations do not originate as coastal sediments – an exception to the rule. Diagram of mineral transformations in the development of acid sulfate soils is depicted below:

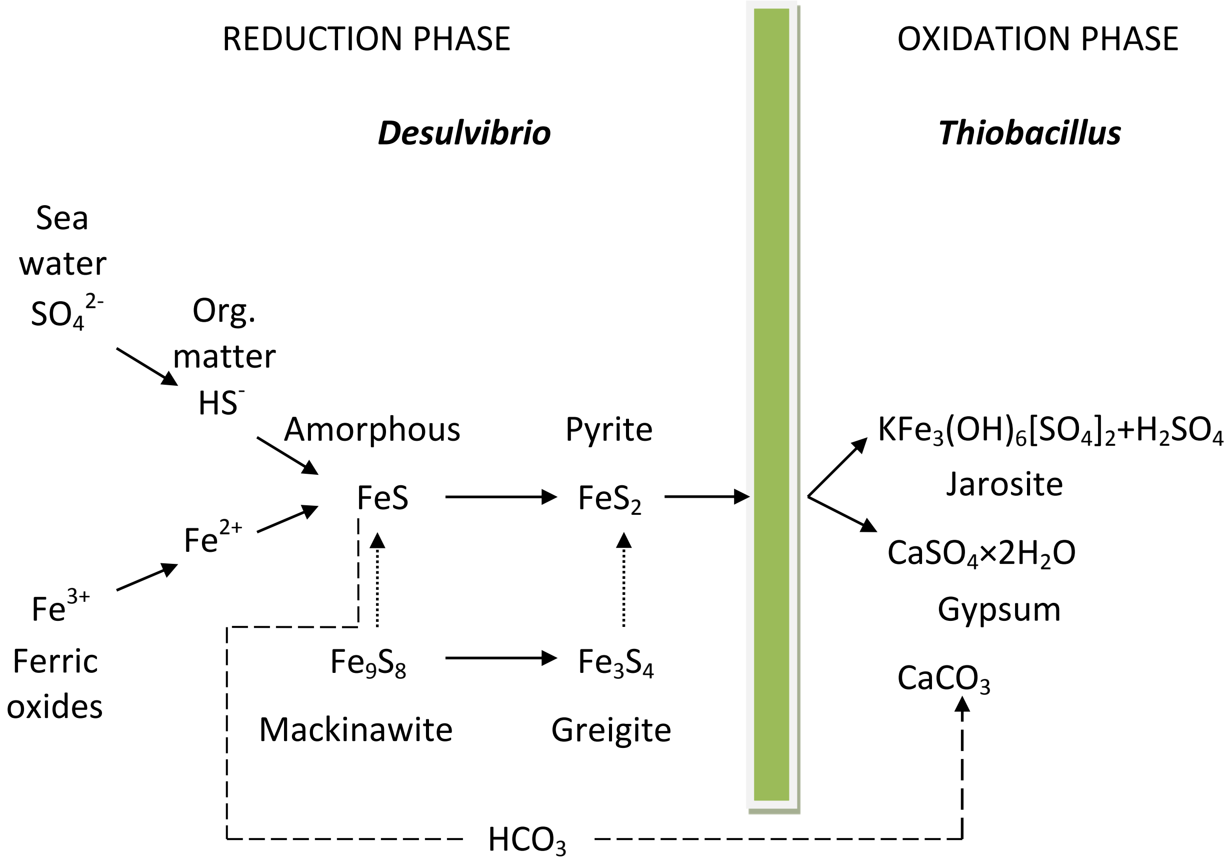


Fig. 43. Mineral Transformations in the Development of Acid Sulfate Soils.

Pyrite (FeS_2) contains 46.5% Fe and 53.5% S. It owes its name to the Greek “πίρ” – fire. Another name is “fool’s gold”. Color: yellow, reminding some dull gold or rather copper; hardness: 6 to 6 ½; specific gravity: 4.9 to 5.2; luster: metallic; streak: greenish or brownish black; fracture: uneven; commonly found in well-formed cubic crystals with striated faces. It may form grainy aggregates, inclusions in igneous and other rocks and earthy masses in ores. In sedimentary rocks, some spherical concretions may be found with a radial internal structure. Cleavage: uneven, sometimes conchoidal. Pyrite is commonly associated with a number of different ores, including copper and gold. Pyrite is a valuable source of sulfur in the manufacture of sulfuric acid. Large areas of pyrite deposits are in the Urals and Altai (Russian Federation).

Chalcopyrite ($CuFeS_2$), known also as copper pyrite, is widely distributed in many rocks and is the principal ore of copper. It contains 34.6 % Cu, 30.5 % Fe and 34.9 % S. It typically occurs in massive forms; its physical properties are: color: brass-yellow or gold (may tarnish green or black); hardness: 3 ½ to 4; specific gravity: 4.2; luster: metallic; streak: greenish black; very brittle. In some localities chalcopyrite may carry gold and also silver. So it is useful to remember that the extreme brittleness of chalcopyrite serves to distinguish it from gold, which is always malleable. The largest deposits of this mineral, more or less close to Ukraine, are those of the Urals, Transcaucasia, Middle Asia and Altai.

Galena (PbS), the most important source of lead, occurs in a wide variety of rocks, including igneous, sedimentary, and metamorphic. It may be found as a replacement in limestone, in veins, or in localized concentrated “pockets”. Copper, zinc, and silver ores are often found associated with galena. Galena has perfect cubic cleavage and its crystals are usually cubic or octahedral. It may also occur in massive, granular, or compact masses. Physical properties: color: lead grey; hardness: 2 ½; specific gravity: 7.4 to 7.6; luster: bright metallic; streak: same as color; fracture: subconchoidal.

As it contains 80 % Pb, it is the most commercially valuable ore, the largest deposits of which lie in the Caucasus, Altai, near the Baikal, Tajikistan and Jakutia. Americans can boast of their galena deposits in the so-called Tri-State Area (Missouri, Kansas, and Oklahoma). There are also commercial quantities in Australia, South America and Europe.

Lead is used in the manufacture of paints (white lead); pipe, shot, solder, metal alloys, and as shielding materials to protect against radioactivity and X-rays.

Sphalerite (ZnS), the primary ore of zinc, is a rather common mineral and its origin and occurrence are similar to that of galena, with which it is commonly associated. It is found in veins in igneous, sedimentary, and metamorphic rocks, and as replacement deposits in limestone. Physical properties: color: yellow, brown, or black; hardness: 3 ½ to 4; specific gravity: 4.1; luster: resinous to adamantine; streak: light to brownish yellow; fracture: conchoidal; cleavage in six directions. Sphalerite contains 67 % Zn. Commercial ores closest to Ukraine are situated in the Northern Caucasus. In the USA, about 18 different states produce zinc, and it is also produced in Canada, Mexico, Peru and Australia.

Zinc is used in galvanizing steel, and in the manufacture of paint, cosmetics, type metal, dry cell batteries, and for a multitude of other purposes.

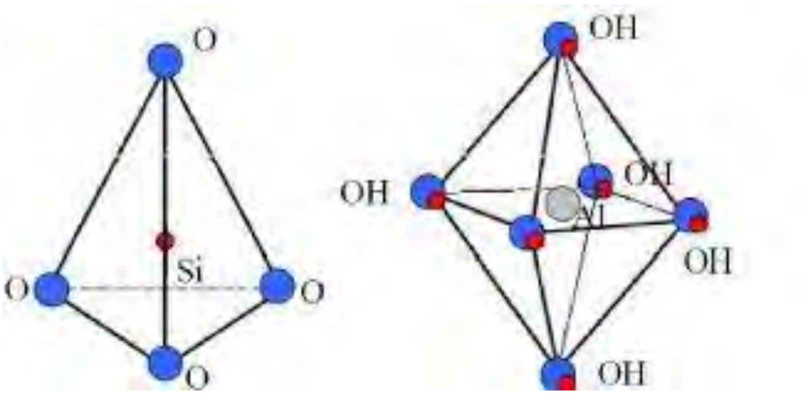
Cinnabar, or mercuric sulfide (HgS) is the most abundant ore of mercury. It contains about 86 % HgS. Its name in Hindi means “red pitch”. Although found in relatively few places, cinnabar occurs in both volcanic and sedimentary rocks, and near hot springs. It occurs most typically in fine-granular or earthy masses. Physical properties: color: bright to brownish red; hardness: 2 to 2 ½; specific gravity: 8.1; luster: adamantine to dull; streak: scarlet: fracture: subconchoidal to uneven.

Native (or free) mercury may also be found in small silvery droplets in certain cinnabar deposits. The origin of cinnabar is mostly low-temperature hydrothermal. In Ukraine, there is a commercially feasible deposit area in Mykytivka (Donbass). The world’s leading producer of mercury is Spain, but Italy is also a relatively large producer. Mercury is used in the amalgamation process of recovering gold and silver from their ores, in the manufacture of explosives, and the manufacture of such measuring instruments as thermometers and barometers.

3.13. Silicates

Silicates are the most widespread minerals in the earth’s crust. They compose 85 % of its weight to the depth of 16 km. Nearly all igneous and metamorphic rocks, as well as the majority of sedimentary ones, are composed of the silicates and quartz. Those also are the most widespread minerals of soils unless the latter were formed in peats and other organic parent materials. Besides quartz, the most abundant primary minerals of soils are the feldspars, micas, pyroxenes, amphiboles, olivine, etc. Minerals of the clay fraction of soils are largely secondary, either inherited from sedimentary rocks or formed by weathering or soil formation. To a lesser extent kaolins, and to a greaser – smectites (montmorillonite), vermiculites, hydrous micas and chlorites are the most important constituents of the solid part of the soil, which determine its absorbing capacity and numerous physical properties important for agronomy.

Silicon is a very abundant cation, which performes a role in the mineral world similar to the role, carbon plays in the organic world. X-ray analysis of the crystal structure of silicate revealed them to be the derivatives of two structural units: a silica tertrahedron and aluminum (and also magnesium, or iron) octahedron:



a b

Fig. 44. Diagram of (a) a silica tetrahedron, and (b) an octahedron of aluminum, magnesium, or iron.

The silicon ion fits in an interstice formed by four oxygen ions. A tetrahedron is formed by the covalent bonding (electron sharing) between oxygen and silicon. The valence (or, rather, charge) of the silicon is +4 and that of the oxygen is -2. Therefore, each tetrahedron (SiO_4^{4-}) has a net charge of -4. The coordination number is the number of ions that can be packed around a central ion. Four oxygen ions can be packed around a silicon ion, resulting in a coordination number of 4. Larger cations have a larger coordination number. Aluminum, iron, and magnesium ions are larger than silicon ions, and they typically have a coordination number of 6. Potassium has a coordination number of 8 and 12, while sodium and calcium – that of 8 (Table 9).

Different silicate minerals are formed, depending on the way the silicon-oxygen tetrahedra are linked together and how the net charge of the tetrahedra is neutralized.

Ions are held together more or less rigidly in a crystal structure, as determined by geometry and by electrical stability. At the same time, some silicates vary widely in chemical composition. Substitution of one element for another in the mineral structure is more the rule than the exception. The terms isomorphic substitution, isomorphism, atomic substitution and solid solution all refer to the substitution of one element for another in ionic crystals without changing the structure of the crystal.

Table 9. Size, Percent Volume in Earth's Crust, Coordination Number and Valence of the Most Abundant Elements in Silicates (H. D. Foth, 1990)

Element	Atomic Radius, Nanometers	Volume Percentage in Earth's Crust	Coordination Number with Oxygen	Valence (Charge)
O	0.132	93.8	–	-2
Si	0.042	0.9	4	+4
Al	0.051	0.5	6 (and 4)	+3
Fe^{+2}	0.083	0.4	6	+2
Fe^{+3}	0.067	0.4	6	+3
Mg	0.066	0.3	6	+2
Na	0.098	1.3	8	+1
Ca	0.099	1.0	8	+2
K	0.133	1.8	8 (and 12)	+1

The more common isomorphic substitutions in silicate structures are Al^{3+} for Si^{4+} in tetrahedral coordination, and Mg^{2+}, Fe^{2+}, or Fe^{3+} for Al^{3+} in octahedral coordination.

Substitutions commonly encerentered in soils produce an excessive negative charge and contribute to the cation exchange capacity of soils.

Depending on the way the silicon-oxygen tetrahedra are linked together, the silicates were subdivided into the following subclasses: single tetrahedral, ditetrahedral and ring structures; single chains, double chains, sheet structures (layer silicates) and framework structures (tectosilicates).

3.13.1. Single-Tetrahedral or Island Silicates (Nesosilicates)

These are the silicates with isolated from each other silicon-oxygen tetrahedra. They are also referred to as island silicates. In olivine, the silicon-oxygen (Si-O) tetrahedra are bonded together by magnesium and/or iron. The crystal is neutral and has the formula $(Mg,Fe)_2SiO_4$. The ratio of magnesium and iron, however, is highly variable, resulting in several mineral species from forsterite Mg_2SiO_4 to fayalite Fe_2SiO_4.

Olivine inherited its name from the olive color. It belongs to rock-forming minerals of ultrabasic and basic extrusive and intrusive rocks (basalts, gabbro, peridotites, etc.). Kimberlite, a peridotite composed of a pyroxene-olivine mixture, is famous for the large number of diamonds that have been extracted from it. Olivine may also form some single-mineral rocks like dunites. Physical properties: color: olive, green, yellowish green, gray-green, brown and even black; streak: colorless (scratch); hardness: 6.5-7.0; specific gravity: 3.3; luster: vitreous; cleavage: nonperfect. The structure of olivine is shown in Fig. 44.

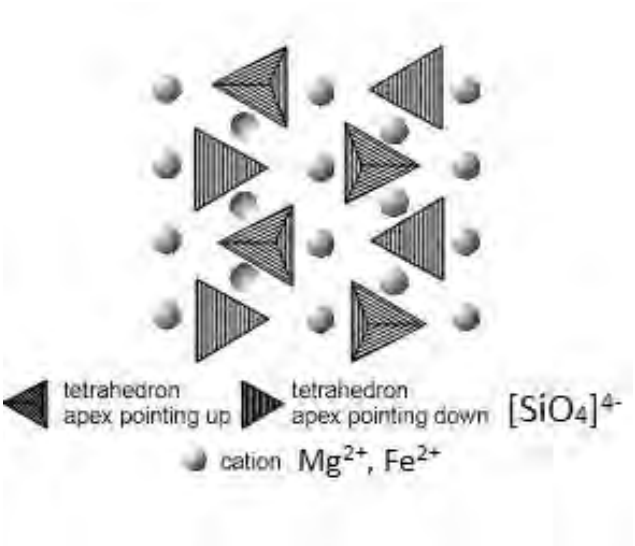


Fig. 45. Single Tetrahedra of Olivine.

Each magnesium ion, or iron ion, is surrounded by six oxygen ions; three oxygen ions form each of the faces of two adjacent Si-O tetrahedrons. The coordination number of magnesium and iron is 6.

The oxygen-silicon bonds are much stronger than the magnesium and iron bonds with oxygen. Consequently, reaction of olivine with water results in the H^+ of the water replacing Mg and Fe ions from the crystal face, rather than the silicon. The weathering of olivine produces serpentine, asbestos, talc, iron oxides, hydromicas, magnesite, etc. In soil environment this mineral weathers rapidly and soils with a high content of olivine may have so much Mg^{2+} in solution as to be detrimental to some plants. Conversely, olivine is absent in many soils because of its low resistance to weathering.

The olivine-containing minerals are widespread in the Urals, Karelia and Eastern Sayan. Low-iron olivine rocks are used as fire-resistant materials. A semiprecious stone chrysolite is employed in jewelry.

3.13.2. Single-Chain Silicates (Inosilicates)

They include a group of pyroxenes (from Greek – fire resistant). If adjacent Si_O tetrahedra share one oxygen, single chains are formed:

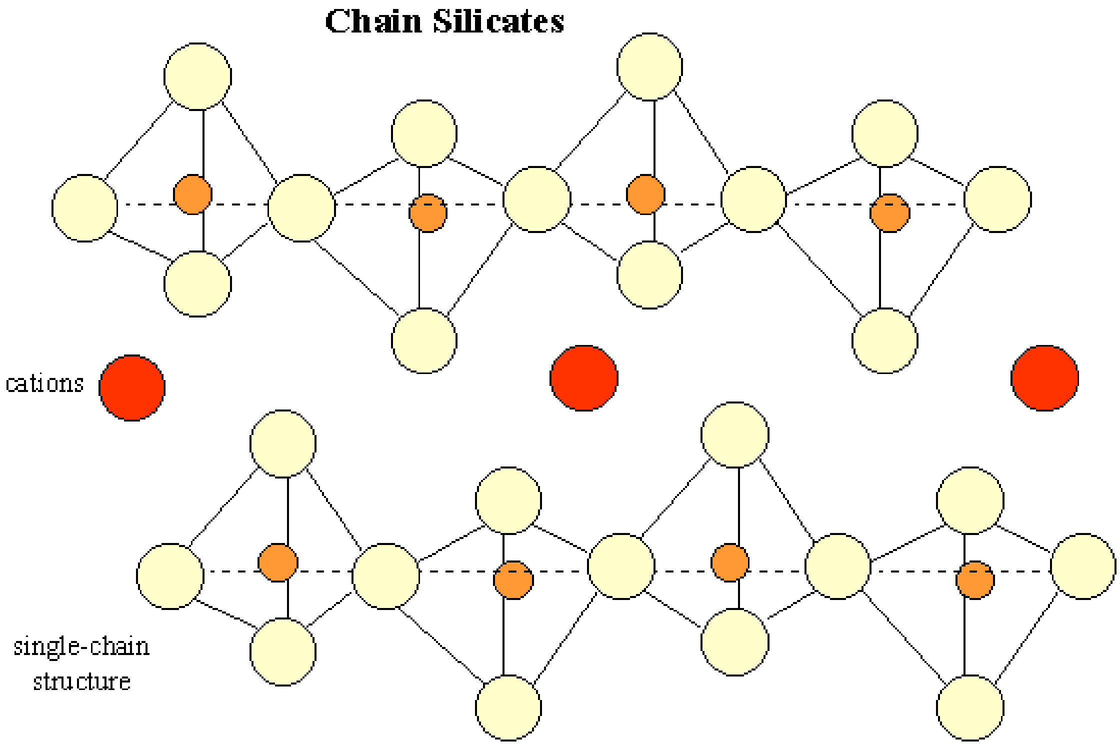


Fig. 46. Structure of a Pyroxene.

The net charge of the chains may be partly or entirely neutralized by the cations of Ca, Mg and Fe. The simplest structured element is represented here by the $[SiO_3]^{2-}$ groups. The bonds between the chains are much weaker than between the tetrahedra within the chains. The minerals, therefore, possess a noticeable cleavage. These minerals are slightly more resistant to weathering than olivine.

Augite $[(Ca, Na)(Mg, Fe, Al)[(Si, Al)_2O_6]$ is a primary mineral composing basic and ultrabasic igneous as well as metamorphic rocks. It is not resistant to weathering and forms talc, kaolinite, limonite, etc. Physical properties: color: black. Greenish black, dark green; hardness: 5-6; specific gravity: 3.3-3.6; luster: vitreous; cleavage: medium; streak: white to grayish green; crystal system: monoclinic.

Enstatite $[Mg_2(Si_2O_6)]$ is also a primary mineral of igneous origin, but unlike augite, it crystallizes in an orthorhombic system. Color: light gray, greenish or yellow-gray; hardness: 5 ½; specific gravity: 3.1-3.3; luster: vitreous; cleavage: two at nearly right angles; streak: white to gray. This mineral is a constituent of some basic and ultrabasic rocks and is encountered in shales. The weathering of enstatite transforms it into a hornblende.

3.13.3. Double-Chain Silicates (Inosilicates)

The minerals are composed of continuous double chains of tetrahedra, sharing alternately two and three oxygens. The elementary structural group is **$[Si_4O_{11}]^{6-}$**.

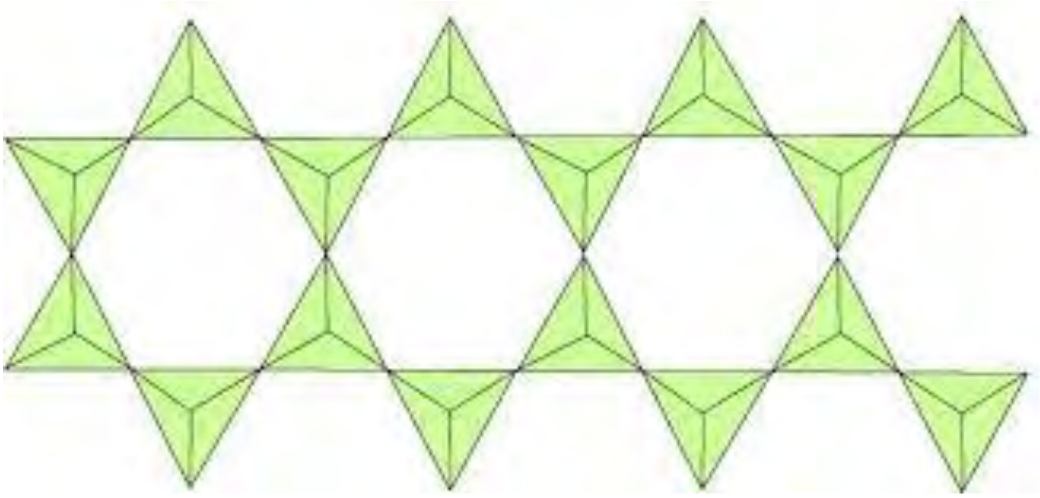


Fig. 47. Amphibole Double Chain.

Like in pyroxenes, the net negative charge of these structures may be neutralized by the metallic ions, primarily Fe^{2+}, Mg^{2+} and some Al^{3+}. Compared to the pyroxenes, the amphiboles are a bit more resistant to weathering. The crystal structures of amphiboles are formed due to a cationic isomorphism; the silicon of the tetrahedra may be substituted by the aluminum. The excessive negative charge is neutralized by the addition of Ca^{2+}, Na^{+} and other cations. As the bond between the parallel bands (pairs of chains) is relatively weaker than between the tetrahedra, the amphiboles display a noticeable cleavage.

The amphiboles are a group of common rock-forming minerals. Because of their great similarity, this group is often confused with the pyroxenes.

Hornblende $[Na, Ca_2(Mg, Fe^{2+})_4(Al, Fe^{3+})(OH,F)_2(Al_2Si_6O_{22})]$ is the most widespread mineral among the amphiboles. Its chemical composition is not constant. The origin is igneous and occasionally – metamorphic. Hornblende is a rock-forming mineral of gabbro, granites, diorites, andesite, etc. It also occurs in loess and other loams as well as in the soils forming in these parent materials. Its physical properties are: crystals: monoclinic prisms, though some amphiboles crystallize in the orthorhombic system; color: black to dark green; hardness: 5 to 6; special gravity 2.9 to 3.3; luster: vitreous (fibrous varieties have silky luster); cleavage: perfect prismatic with angles of 56 degrees and 124 degrees. The weathering of hornblende yields carbonates, opal, iron hydroxides, etc.

Isomorphous admixtures such as the trace elements are very important in soil environments.

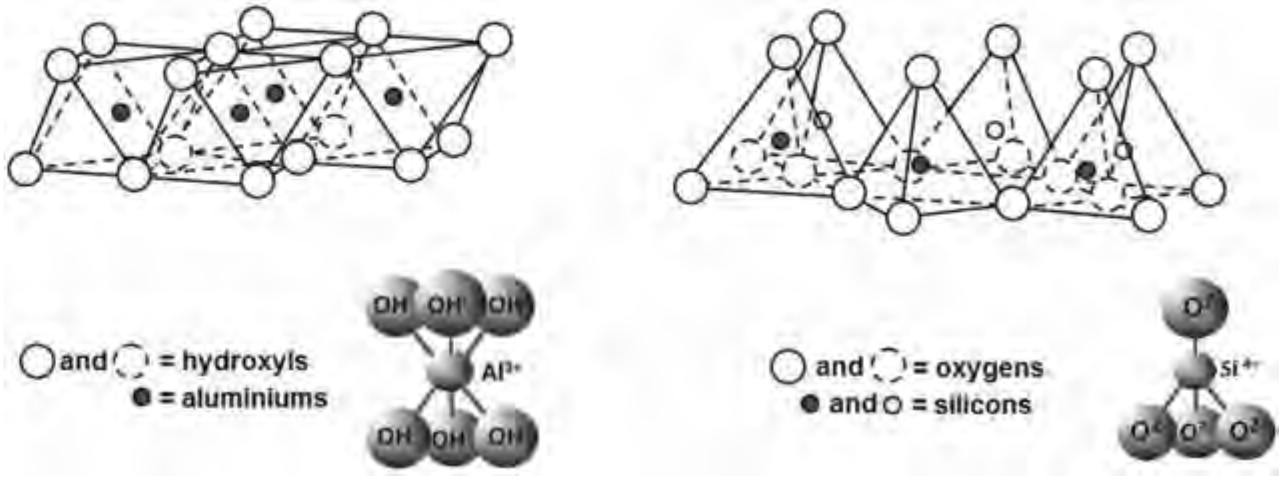
Actinolite and tremolite are other interesting amphiboles. Some mineralogists treat them as a single mineral. They commonly occur in long prismatic, bladed crystals, or as fibrous or as bestiform (resembling asbestos) masses. Fibrous tremolite is used to some extent as asbestos in fireproofing and insulation. Ferromagnesial mineral weathering may be represented by the following generalized scheme:

$$(\text{Olivines, pyroxenes, amphiboles})+O_2+H^+\rightarrow\underset{\text{(precipitated)}}{Fe(OH)_3}+\underbrace{Mg^{2+}+Ca^{2+}+\text{silica residue}}_{\text{(leached)}}$$

Soils, dominated by these minerals in the fine silt and clay fractions, are the minimally weathered soils all over the world, but are mainly the soils of the desert regions where limited water keeps chemical weathering to a minimum. Minimally weathered soils are generally fertile, as they still contain some nonleached nutrient elements.

3.13.4. Layer (Sheet) Silicates (Phyllosilicates)

Although the common primary minerals include island, chain, sheet, and framework silicates, the most stable and persistent silicates, which occur as weathering products in the clay fraction of soils, are sheet (or layer) silicates. A typical layer silicate may be a combination of two types of structural units. One is a two-dimensional array of Si-O tetrahedra (the tetrahedral sheet). The second is a two-dimensional array of cations (generally Al^{3+}, Mg^{2+}, Fe^{2+}, or Fe^{3+}) in octahedral coordination with oxygen or hydroxyl anions (the octahedral sheet). The tetrahedral and octahedral sheets are bonded together by a layer of shared oxygens.



A B

Fig. 48. Structure of a metal-hydroxyl octahedron (A) and tetrahedral sheet (B).

The simplest structural group $(Si_4O_{10})^{4-}$ carries a negative charge neutralized by metallic cations (Mg^{2+} and Al^{3+} being the most typical).

Layer silicate minerals are differentiated by (1) the number and sequence of tetrahedral and octahedral sheets, (2) the layer charge per unit cell of structure; (3) the type of interlayer bond and interlayer cations, (4) the cations in the octahedral sheet, and some other peculiarities. Numerous structured combinations of the tetrahedral sheet with octahedrally coordinated metal cations are possible.

The bands of metal-hydroxyl octahedral form sheet-like layers which alternate with the layer of Si-O tetrahedra. The sheets (or nets), in their turn, alternate and unite each other into packets through the ions of Mg^{2+}, Al^{3+}, K^+, Na^+, Ca^{2+} and OH^-. Structural anionic radicals in these packets are represented by Si-O groups, for example, $[Si_2O_5]^{2-}$ or $[Si_4O_{10}]^{4-}$.

Depending on the number and sequence of sheets, there are two-, three- and four-layer packets. Such internal structure of layer silicates is reflected in their specific properties: more or less perfect cleavage in one direction, parallel to the plane of sheets, relatively low hardness, tabular or needle-like appearance of crystals. Such platy structure is well noticeable in micas, but one cannot easily observe it in the other layer silicates.

A part of silicon ions of both tetrahedral and octahedral “nets” may be substituted by aluminum ions at the expense of a well developed cationic isomorphism. That is why a part of the minerals, belonging to this group, are called aluminosilicates. Al^{3+}, in its turn, may be substituted by Fe^{2+}, Fe^{3+}, Cr^{3+}, Mn^{2+}, etc. and Mg^{2+} may be substituted by Fe^{2+}, Co^{2+} and other cations.

Excessive negative charge contributed to by isomorphic substitutions as well as fine dispersion (small size of particles) especially of clay minerals of imparts to these layer silicates a considerable cation exchange capacity.

Different groups of minerals belong to the layer silicates, for example, talc, serpentine, clay minerals, micas, hydromicas, etc. But all this variety of layer silicates may be divided into two types: (1) the silicates without developed isomorphism and (2) the silicates with developed isomorphism.

A. Silicates without developed isomorphism. They include the subgroups of two- and three-layer silicates.

Two-layer silicates. A two-layer packet is made up with a layer of tetrahedra and a layer of octahedra. Due to a close distance between the packets (c-spacing, 0.72 nm), these minerals possess a poor water-imbibing capacity. Within the octahedra there are the cations of magnesium or aluminum. The bonding between the neighboring packets is weak, which causes perfect cleavage and low hardness of two-layer silicates. The most widespread representatives of the minerals are serpentine and kaolinite (a clay mineral) (see Fig. 49).

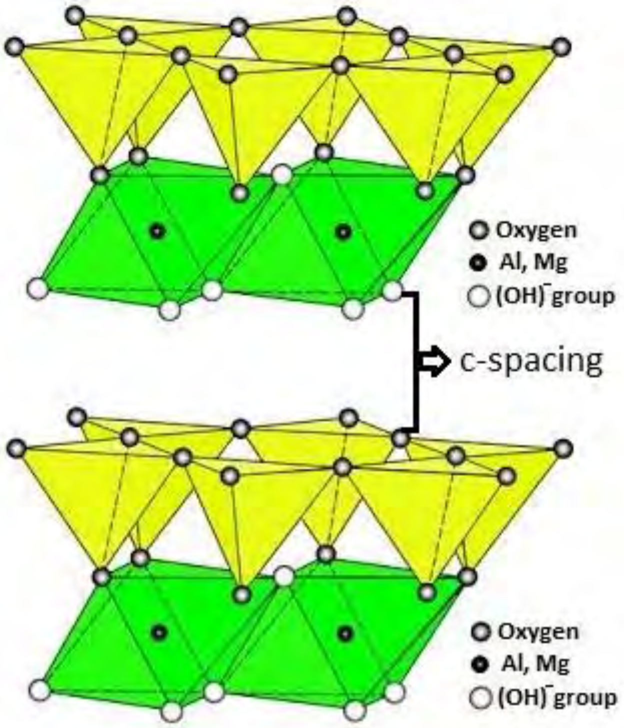


Fig. 49. Schematic Diagram of Tetrahedral (yellow) and Octahedral (green) Layers in the Packets of Layer Silicates.

Serpentine (asbestos) – $Mg_6(OH)_8[Si_4O_{10}]$ is a representative of group of hydrous silicates, which commonly occur in compact masses which feel smooth or greasy. Common, or massive, serpentine is actually a weathered or metamorphosed peridotite. Spots of green olivine alternate with those of gray-green serpentine. Luster: greasy or resinous; hardness: 2 ½ to 4; specific gravity: 2.5 to 2.8; streak: white; cleavage: conchoidal to splintery fracture.

Chrysotile, a fibrous variety of serpentine, is the principal mineral used as asbestos, a very important fire-proof material. There is evidence that contacts with asbestos are dangerous to human health because of its being cancerogenic.

Kaolinite $[Al_4(OH)_8(Si_4O_{10})]$ is of the simplest structure among the clay minerals. Kaolin, formed primarily by the decomposition of rocks containing large amounts of feldspar, is an impure mixture of a variety of clay minerals. The structural unit of the kaolin group is formed by the superposition of a tetrahedral sheet upon an octahedral sheet (Fig. 50).

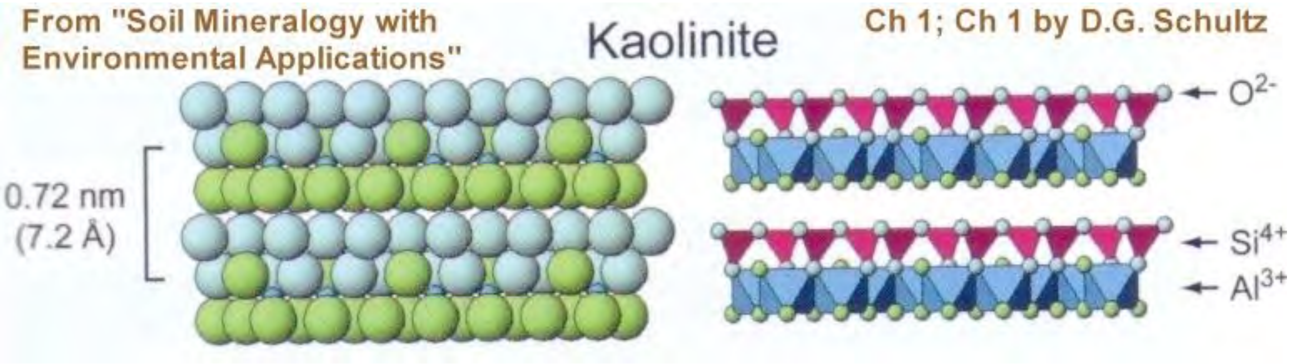


Fig. 50. Schematic structure of kaolinite.

Such minerals are referred to as 1:1 layer silicates. The apical oxygen of the tetrahedral sheet are shared by the octahedral sheet, forming a common plane of oxygen ions within the structure. In the shared plane, two-thirds of the oxygen ions are shared between Si^{4+} and Al^{3+}. The remaining one-third of the oxygen ions have their charge satisfied by H^+ to form OH^- groups. The upper surface of kaolin is a layer of closely packed OH groups, but the bottom surface is composed of hexagonally open-packed oxygens and OH groups recessed within hexagonal (ditrigonal) openings.

Occurring in soft, compact, earthy masses, kaolinite is characterized by its dull earthy luster, greasy feel, and rather distinctive property of adhering to the tongue. It becomes plastic when wet and has an earthy or claylike odor when breathed upon. Hardness ranges from 1 to 2 ½; specific gravity from 2.2 to 2.6. It is widely used in the ceramic and many other industries.

Kaolinite occurs commonly in soils, often as hexagonal crystals with an effective diameter of 0.2 to 2 nm. Halloysite is a form of kaolinite in which water is held between structural units in the basal plane, yielding a c-spacing of 1.0 nm when fully hydrated (see Fig. 51).

Most kaolin structural units, however, are held together in the basal plane by hydrogen-bonding between oxygen ions of the tetrahedral sheet and hydroxyl ions of the octahedral sheet. Surface area is limited to external surfaces and hence is relatively small (10 to 20 $m^2 * g^{-1}$). Kaolinite is a coarse clay with low colloidal activity, including low plasticity and cohesion, and low swelling and shrinkage. Most of the 10 to 100 $mmoles * kg^{-1}$ cation exchange capacity (CEC) of kaolinite has been attributed to dissociation of OH^- groups on clay edges. The CEC of kaolinite is highly pH-dependent (grows with growing pH).

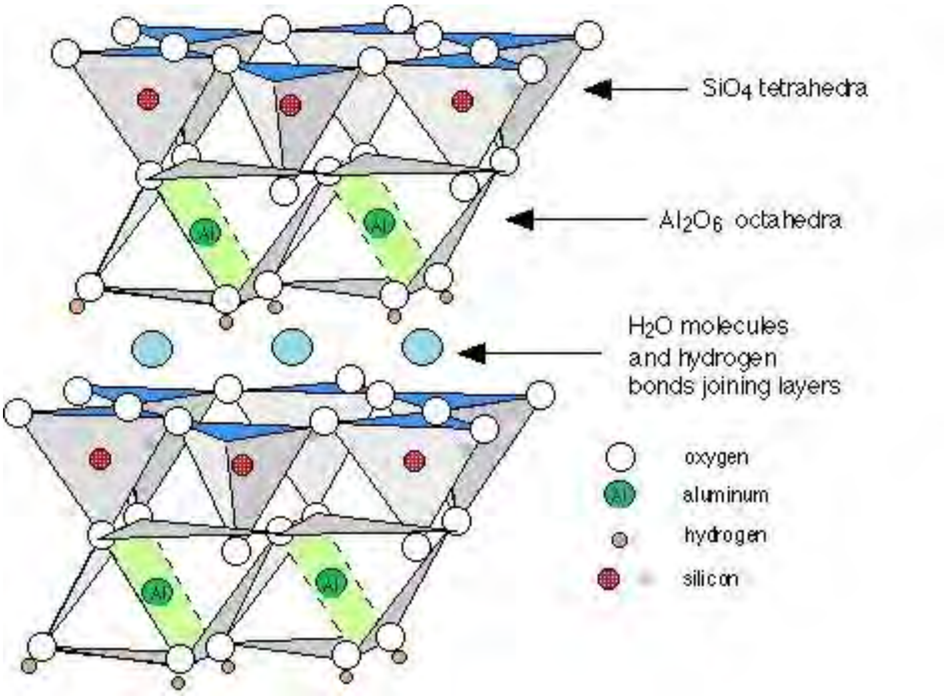


Fig. 51. Structure of Halloysite.

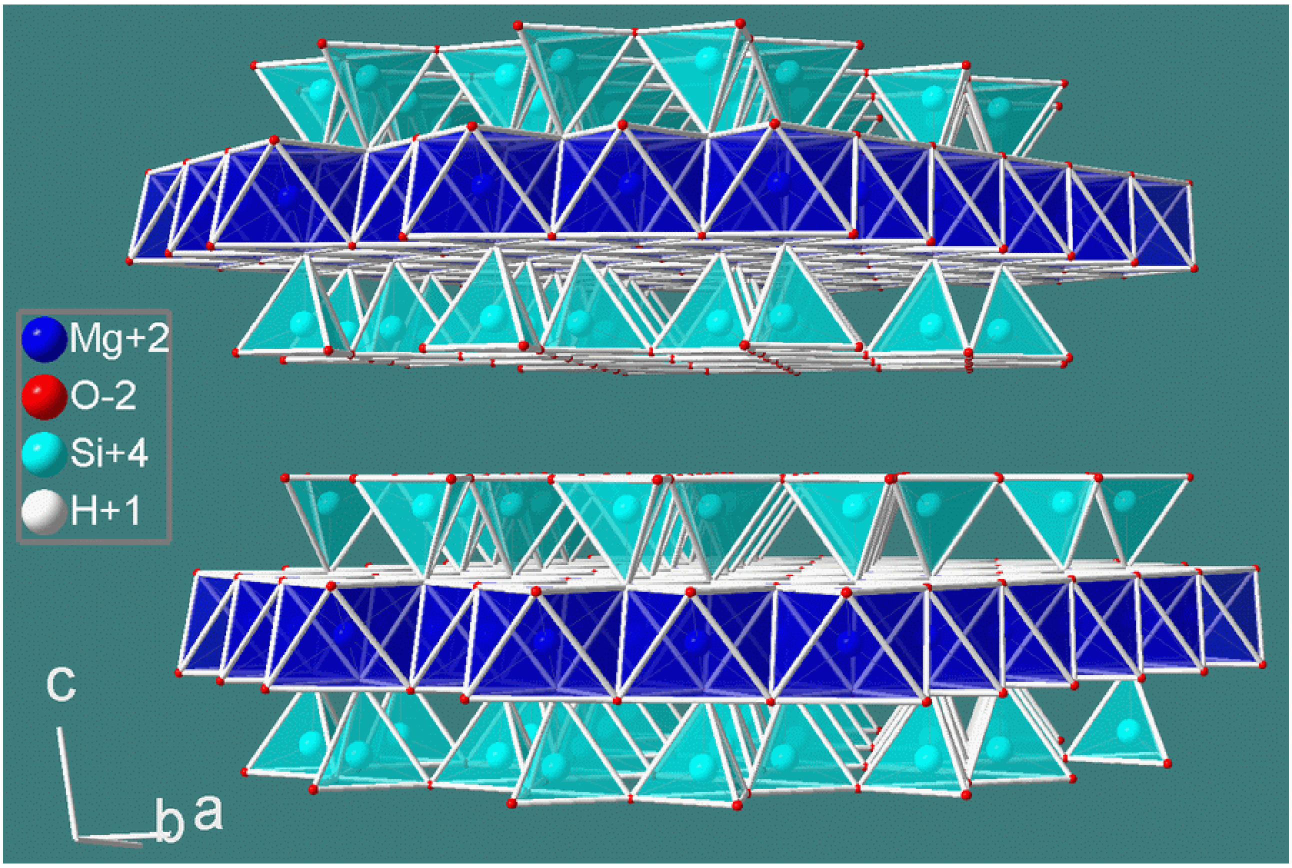


Fig. 52. Structure of Talc (from Wikipedia).

Three-layer silicates have their structural units (packets) composed of three sheets: between the two tetrahedral sheets there is sandwiched a single octahedral sheet. They are also referred to as the 2:1 layer silicates. These minerals possess a perfect cleavage easily disintegrating into plates and scales.

Talc $\{Mg_3(OH)_2[Si_4O_{10}]\}$ forms leafy, scaly, fibrous and often compact masses. Color: green, light-grey, brown; streak: white; luster: vitreous, greasy or pearly; hardness: 1; specific gravity: 2.7-2.8. This mineral forms in the upper strata of the earth's crust by the weathering of ultrabasic and basic igneous rocks rich in olivine:

$$\underset{\text{olivine}}{4(MgFe)[SiO_4]} + H_2O + Co_2 + O_2 = \underset{\text{talc}}{Mg_3(OH)_2[Si_4O_{10}]} + MgCo_3 + 2Fe_2O_3$$

Talc is used in paper, rubber, leather and other industries, for the production of fire-resistant utensils and bricks as well as in the insulators. Large deposits are known in the Urals.

B. Silicates with Developed Isomorphism.

Structural element formed by isomorphic substitution of cations in the tetrahedra is $[AlSi_3O_{10}]^{5-}$. A cation needed for the compensation of excessive negative charge is located between the structural units (packets). Isomorphic substitutions within the octahedra also beget a negative charge but of a smaller order. This affects the stability of structural units.

Depending on the extent of a negative charge compensation in the octahedra and interpacket spacings a several groups of silicates have been identified: micas, hydromicas, vermiculites, chlorites, etc.

Micas are very widespread rock-forming minerals characterized by perfect basal cleavage (micaceous cleavage).

Muscovite $[KAl_2(OH)_2(AlSi_3O_{10})]$, like all micas, is easily identified as it typically occurs in paper-thin, shiny, elastic cleavage plates. This mineral, known also as white mica or "isinglass", is usually transparent. It is colorless, gray, or light brown, has a hardness of 2 to 2 ½, specific gravity of 2.8 to 3.1, and a pearly to vitreous luster. Occurring typically in thin, elastic, scale-like crystals, muscovite is a common constituent of certain granites and pegmatites. It also occurs in certain metamorphic and sedimentary rocks.

Commercially, muscovite is used in manufacturing electrical equipment, insulating cloth and tape, lamp chimneys, lubricants, paints, and Christmas tree "snow". Muscovite contains 11.8 % K_2O. Its crystal structure is shown in Fig. 53.

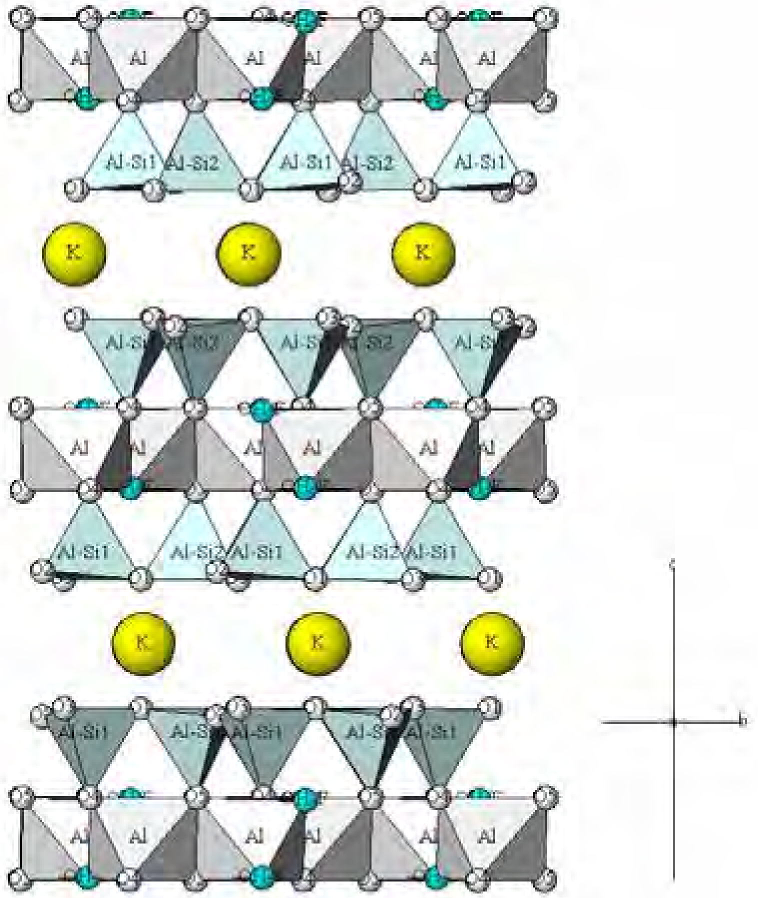


Fig. 53. Schematic Structure of Muscovite (from staff.aist.go.jp/nomura-k/english/itscgallary-e.htm).

Biotite $[K(Mg, Fe_3(OH, F)_2AlSi_3O_{10}]$, or black mica, is a very common mica and commonly occurs in association with muscovite. It is found in many igneous and metamorphic rocks, where it is seen as thin, platy, shiny black sheets or scales. Biotite is typically dark brown to black (sometimes green). The physical properties of biotite are essentially the same as those of muscovite. Unlike muscovite, biotite mica has very little commercial value.

Micas are abundant in soils, largely a primary minerals inherited from parent materials. They are precursors for other 2:1 layer silicates, notably vermiculites. Altered mica containing less K^+ and more water than well-ordered mica is called hydrous mica (formerly illite). Interlayer K^+ is so strongly adsorbed that it is not exchanged in CEC determination: the CEC is only 200 to 400 mmoles (+) kg^{-1}. Total surface area is about 70 to 120 $m^2 * g^{-1}$. Micas are nonswelling and are only moderately plastic. “Fixed” K^+ is released slowly during weathering.

Vermiculite {$(Mg, Fe^{2+}, Fe^{3+})(OH)_2[(Si, Al)_4O_{10}]*4H_2O$} contains up to 5 % K_2O and is a widespread hydromica. Color: brown, yellowish brown, gold-yellow; crystal system: monoclinic; luster: greasy; specific gravity: 2.4-2.7; when heated-swells 15-20 times the initial volume.

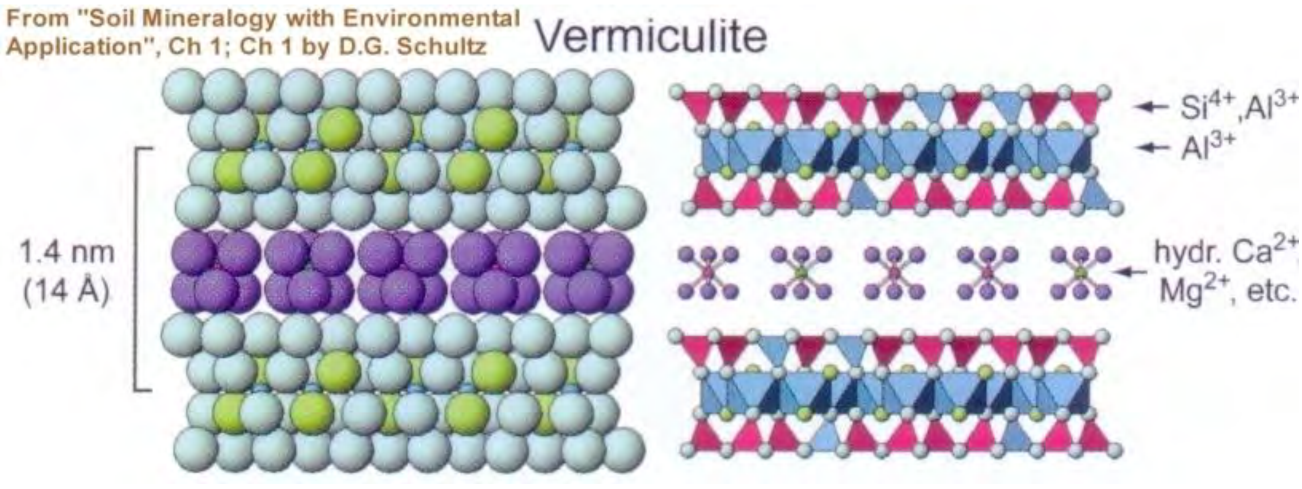

From "Soil Mineralogy with Environmental Application", Ch 1; Ch 1 by D.G. Schultz

Fig. 54. Schematic Structure of Vermiculite: the space between the structural units is filled with cations and water.

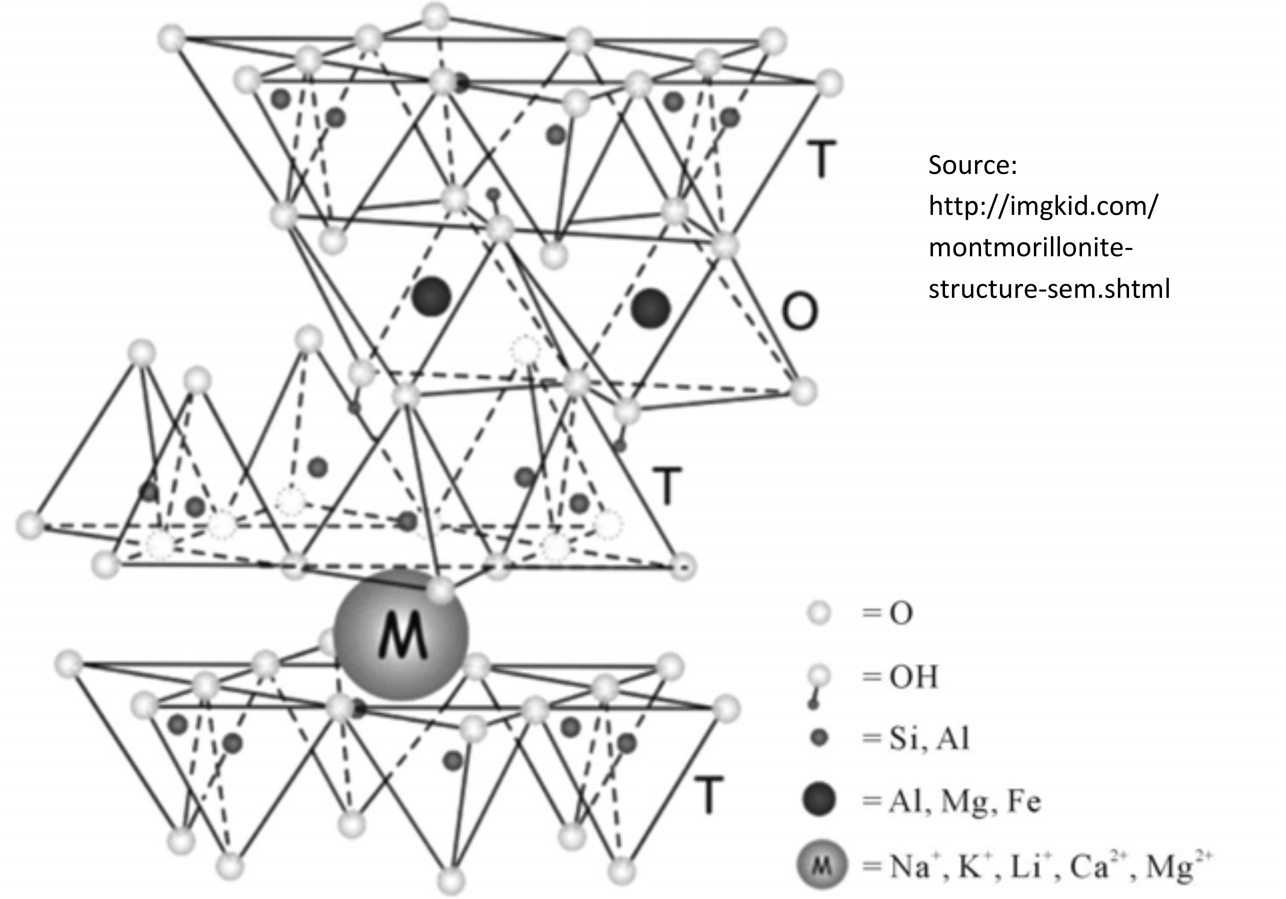
Weathering of the precursor micas has replaced the interlayer K^+ by (predominantly) exchangeable Mg^{2+} and has expanded the c-spacing to 1.4 to 1.5 nm in most cases. Cation exchange capacity: 1200 to 1500 mmoles (+) kg^{-1}, and only slightly pH-dependent. Vermiculite swells less than montmorillonite because of its higher layer charge. Total surface area, when not K^+ or NH_4^+ saturated, ranges from 600 to 800 $m^2 * g^{-1}$.

Chlorites. An idealized half unit-cell formula is $(AlMg_2(OH)_6)_x(Mg_3(Si_{4x}Al_x)O_{10}(OH)_2$. They occur extensively in soils and are examples of 2:1:1 layer silicates. These minerals resevuble the micas and commonly occur in foliated or scaly masses, although tabular six-sided crystals may occur. The chlorite minerals are typically green, have one perfect cleavage, a hardness of 1 to 2 ½, specific gravity of 2.6 to 3, and leave a greenish streak. They are common constituents of many igneous and metamorphic rocks. CEC ranges from 100 to 400 $mmoles*kg^{-1}$, and total surface area values from 70 to 150 m^2*g^{-1}. Chlorite occurs commonly in sedimentary rocks and in productive soils derived therefrom. Serpentine-derived soils contain chlorite and often are infertile because of their high magnesium and low calcium contents.

<u>Glauconite</u> $\{K(Fe^{3+}, Al, Fe^{2+}Mg)_2(OH)_2[AlSi_3O_{10}]*nH_2O\}$ – a secondary aluminosilicates of prehistoric sea chemical origin, common in sedimentary rocks. In Ukraine, green glauconite sands compose Kharkiv Suite (Oligocene epoch, Tertiary period). Glauconite contains 2-8 % K_2O and is used for the manufacturing of green paint and to mellow the hardness of water.

<u>Montmorillonite (Smectites)</u> $\{(Ca, Na)(Mg, Al, Fe)_2(OH)_2(Si, Al)_4O_{10}*nH_2O\}$ is a 2:1 layer silicate, freely expanding because of the relatively low layer charge. The c-spacing varies with the exchangeable cation and the degree of interlayer salvation (from 0.95 to 1.5 nm or more). Complete drying yields a spacing of 0.95 to 1.0 nm, and full hydration can swell the layer to a distance up to tens of nanometers.

Montmorillonite is formed from the weathering of granites and diorites. Specific gravity: 2-2.5; color: light green, white or blue.



Source:
http://imgkid.com/
montmorillonite-
structure-sem.shtml

Fig. 55. Montmorillonite pictured in terms of its tetrahedral (T) and octahedral (O) sheets. M is an exchangeable cation connecting sheets of silica/oxygen tetrahedrons (After Güven). Water molecules are associated with the cation and form up to three water layers.

CEC for montmorillonite ranges from 800 to 1200 mmoles*kg^{-1}, being only slightly pH-dependent. Total surface area is from 600 to 800 m^2*g^{-1}. Montmorillonite has high colloidal activity, high plasticity and cohesion. Normally occurs as a fine clay with irregular crystals having an effective diameter of 0.01 to 1 mm. Smectites are common in Vertisols (the soils developed in expanding clays) and in soils of alluvial plains.

3.13.5. Framework Silicates (Tectosilicates)

These silicates' structure is a continuous framework of tetrahedra each sharing all four oxygens. Thus the tetrahedra do not have free valencies and the addition of cations to them is impossible. But a part of silicon ions in this structure are substituted by the aluminum ions. Thus an excessive negative charge appears which is compensated by the cations of potassium, sodium or calcium. The basic structural group of the framework silicates is $\mathbf{(Si, Al)_4O_8}$. Feldspars, the most important group of rock-forming minerals, belong to the framework silicates. They are so very abundant, that they have been estimated to make up as much as 60% of the earth's crust. Feldspars are found in almost all igneous rocks as well as in many sedimentary and metamorphic rocks. The group of feldspars has been divided into two principal subgroups (1) orthoclases and (2) plagioclases.

Orthoclase $K[AlSi_3O_8]$ is a rather common potash feldspar. It is transparent to translucent and in color may range from colorless to white, gray, flesh-red, yellow, pink and even greenish-blue. It has a vitreous luster, hardness of 6, and specific gravity of 2.5 to 2.6. It crystallizes in the monoclinic system and has two easy cleavages making an angle of 90 degrees. Fracture is uneven to somewhat conchoidal; streak is white or uncolored. Orthoclase can usually be distinguished from plagioclase feldspar by absence of striations (stripes). Microcline is another potassium aluminum silicate of the same chemical composition as orthoclase. However, they crystallize in different crystal systems and differ in certain other physical characteristics. The cleavage in microcline make an angle of 85 degrees (hence the name – "slightly deviating").

Plagioclases. These are the Isomorphous mixture series of two minerals: albite $\mathbf{Na[AlSi_3O_8]}$ and anorthite $\mathbf{Ca[Al_2Si_2O_8]}$. They are also popularly known as soda-lime feldspars and are common in many igneous rocks and certain metamorphic rocks. In the earth's crust they are represented by (% of albite is given in brackets): albite (100-90%), oligoclase (90-70%), andesine (70-50%), labrador (50-30%), bytownite (30-10%) and anorthite (10-0%). Albite is an acid plagioclase, andesite and labrador-neutral or medium ones, while anorthite – a basic one. Light colored acid plagioclases in the series from albite to anorthite gradually become darker and heavier in specific gravity. So the colors are white, yellow, reddish gray, gray to black; luster: vitreous; hardness: 6; specific gravity 2.6 to 2.8; transparent to translucent; two good cleavages at almost 90 degrees to each other and two poor prismatic cleavages.

Albite $NaAlSi_3O_8$ occurs in tabular crystal form; doubles are encountered occasionally; cleavage: perfect; color: white, colorless, bluish, brownish or reddish; streak: white; luster: vitreous; hardness: 6-6 ½; specific gravity: 2.6. As a rock-forming mineral occurs in granites, pegmatites and crystal shales. Weathering transforms it into a Kaolin. Deposits in pegmatitic veins are encountered in the Urals.

Oligoclase (Na, Ca) $[Al_{1-2}Si_{3-2}O_8]$ has a perfect cleavage; color: gray, greenish, yellow, brown or reddish; semitransparent variety goes under the name of moonstone and is characterized by a play of colors called opalescence; streak: white; luster: metallic; hardness: 6-6 ½; specific gravity: 2.6. Is encountered in pegmatites, granites and syenites.

Labradorite (Na, Ca)$[Al_{1-2}Si_{3-2}O_8]$ crystallizes in triclinic system forming tabular crystals or compact grainy masses and sometimes – doubles. Cleavage: perfect; color: white or gray with white to bluish and green internal flashes (a play of colors on the thin planes of doubling crystals); hardness: 6-6 ½, specific gravity: 2.0-2.7; composes basic igneous and metamorphic rocks (labradorite, gabbro, basalt and diabase) and sometimes – medium (neutral) igneous rocks (diorite and andesite). The deposits of labrador (labradorite) are encountered in Zhitomir and Volyn regions of Ukraine as well as on the Labrador Peninsula.

Anorthite $Ca[Al_2Si_2O_8]$, a basic plagioclase, forms prismatic and short columnar crystals and compact grainy masses. Cleavage: perfect; color: from white to a varying degree of gray; luster: vitreous; hardness: 6-6 ½; specific gravity: 2.7; is encountered in intrusive and volcanic rocks in combination with limestones.

All feldspars are of considerable economic importance. Orthoclase is used in the manufacture of china, porcelain and scouring powders. Feldspars are also used in making paints, enamels, and glass. Plagioclase feldspars are less commonly used then potash feldspars, but some are used in the ceramics industry and as beautiful decorative stones.

3.14. Stability of Parent Material Minerals

The major minerals of igneous rocks are, in decreasing order of general abundance: feldspars, quartz, and biotite and muscovite micas. Other minerals in igneous and sedimentary rocks are generally present in lesser amounts. Granitic or acid (>66% SiO_2) igneous rocks are richer in silicon and potassium and poorer in magnesium and iron, than basaltic or basic igneous (45 to 52% SiO_2) rocks.

Sedimentary rock materials have already passed through some weathering before the rock is formed. Their composition represents depletion of weatherable elements. The resistance of igneous minerals to weathering is the same as the order of crystallization from cooling magmas (Fig. 56).

Increasing stability to weathering

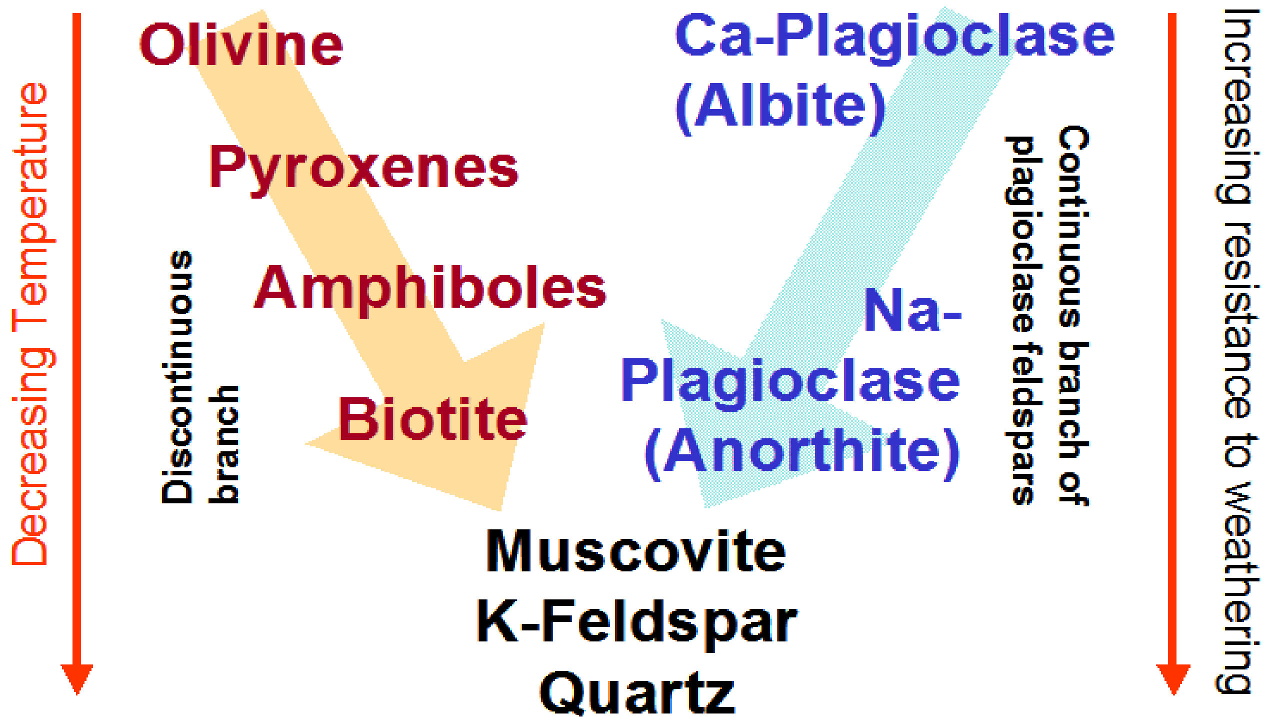


Fig. 56. Stability to weathering of some minerals in igneous and metamorphic rocks (Adapted from S. S. Goldich, J. Geology, 46:38 (1938).)

Minerals that are most stable at high temperatures are the least stable at low temperatures. The weathering rate generally increases with increasing content of alkali and alkaline earth cations. A factor affecting mineral weatherability is the degree to which the tetrahedra are linked together. Feldspars and quartz are three-dimensional networks of tetrahedra and they are very stable. Hence, quartz is very persistent in soils. Feldspar would also be resistant to weathering if not for the other factors which outweigh this structural effect. Smaller particles weather more rapidly, but the size effect is great only when the particles are less than several micrometers in size. Cleavage planes allow particles to be more easily broken apart. Feldspars and micas have clearly defined cleavage planes which hasten the rate of mineral breakdown.

3.15. Mineral formation in Soils (from H. L. Bohn, B. L. McNeal and G. A. O'Connor. Soil Chemistry. 2nd Edition. N. Y. 1985)

The formation of secondary minerals in soils generally results from the combination and addition of ions and molecules from the soil solution to the solid phase. Only relatively recently has the slow kinetics of such reactions been appreciated. Secondary minerals can precipitate from solutions containing the proper constituent ions and $Si(OH)_4$. Despite many reservations, the thermodynamic data now available provide a reasonably satisfactory picture of solution conditions under which well-defined minerals form and decompose during weathering and soil development. The equilibrium between Kaolinite and gibbsite can be written:

$$\underset{\text{kaolinite}}{Al_2Si_2O_5(OH)_4} + 5H_2O = \underset{\text{gibbsite}}{2Al(OH)_3} + \underset{\text{soluble silica}}{2Si(OH)_4}$$

The equilibrium constant for the reaction is $K = (Si(OH)_4)^2 = 10^{-8.4}$. The equilibrium constant defines the soluble silica activity, or concentration, at which gibbsite and Kaolinite are in equilibrium: $(Si(OH)_4) = (10^{-8.4})^{1/2} = 10^{-4.2}$.

At $Si(OH)_4$ activities less than $10^{-4.2}$, gibbsite is stable and Kaolinite will not form. Any Kaolinite present will decompose to gibbsite and soluble $Si(OH)_4$. At $Si(OH)_4$ activities greater than $10^{-4.2}$, Kaolinite is the stable solid. Gibbsite is unstable at these concentrations and will react with $Si(OH)_4$ to form Kaolinite.

Ions in soils, however, are in constant flux because of plant and microbial uptake and subsequent organic decay. This continual input of fresh amorphous material presents attainment of equilibrium.

Carbonates (particularly $CaCO_3$) accumulate in soils in the regions of limited rainfall. But however varied are the environmental conditions, carbonate accumulation can be represented by the equation: $Ca^{2+} + H_2O + CO_2 = CaCO_3 + 2H^+$. Increasing P_{CO2} causes $CaCO_3$ to react further: $CaCo_3 + CO_2 + H_2O = Ca^{2+} + 2HCO_3^-$ so that $CaCO_3$ redissolves with increasing CO_2 concentration in the gaseous phase.

The solubility product of $CaCO_3$ in equilibrium with water and gaseous CO_2 yield an acidity of $(H^+) = (10^{-13.5\ PCO2})^{1/2}$. $pH = 6.75 – 0.5 \log P_{CO2}$.

The pH of $CaCO_3$ solution in soil environment: $pH = 6.75 + ½pCO_2$. The pH of any carbonate soil is thus within 7.0 and 8.5 (theoretically 8.4) if exchangeable sodium is low. Where drainage waters from surrounding soils accumulate and where water lost by percolation is small compared to the amount of water evaporated, soluble salts (evaporites) tend to accumulate.

As soils weather, soil colloids become enriched in oxides, hydroxides and hydroxyoxides of Al, Fe and Ti, relative to silicon. These materials form allophane and zeolites. Allophane is a general name for amorphous aluminosilicates gels.

Zeolites are three-dimensional framework structures, like the feldspars. Their tetrahedra may be linked into 4-, 6-, 8-, or 12-membered rings, joined together less compactly than in the fieldspars.

Allophane may have a high cation exchange capacity in neutral to mildly alkaline solutions (of the order of 1500 mmoles*kg^{-1}).

Questions and Assignments

1. Define a mineral. Name the essential groups of primary and secondary minerals.
2. Name the most widespread rock-forming minerals.
3. What do you know about physical weathering?
4. Give examples of oxidation, hydration, and carbonation in chemical weathering.
5. List the essential classes of minerals. On what principles had the classification been built?
6. Physical Properties of Minerals.
7. Describe native elements and their importance in industry.
8. Which metallic or ore minerals do you know and to which classes do they belong?
9. Describe igneous processes of mineral formations.
10. Phosphates and their role in Agronomy.
11. Carbonates and their role in Agronomy.
12. Clay minerals and soil properties.
13. Weathering resistance of various silicates.
14. Structural classification of silicates.
15. Surface area and CEC of Kaolinite and smectites. What properties do these minerals impart to soils?
16. Compare kaolinite, smectites, and chlorite. What common and what different properties and why do these minerals possess?
17. Characterize the stability of primary minerals in soil environment.
18. What pH values can be expected in carbonatic (calcareous) soil solution? How can it affect plant nutrition?

Chapter 4

Introduction to Petrology

4.1. Rocks, the subject-matter of petrology

Petrology, one of the basic geologic subsciences, is the study of rocks – their mineralogy, origin and evolution. In the simplest sense, rocks can be defined as aggregates of minerals. Although monomineralic rocks exist (e.g., limestone and quartzite), most rocks consist of several minerals. They are called polymineralic. Granite, a very common polymineralic rock, consists of feldspar, quartz, hornblende, mica and other minerals.

About 1000 rocks are known, which according to their origin are divided into **three groups**: (1) ***igneous***, (2) ***sedimentary***, and (3) ***metamorphic***. These three groups compose the earth's crust rather unevenly. Igneous rocks occupy 95% of the weight of the crust to the depth of 16 km. The volume of sedimentary rocks does not exceed 3% of the lithosphere, but they cover about 75% of the earth's surface. The metamorphic rocks are the least widespread but may be economically very important. Many of the challenges we, as civilization, currently face, and must confront, directly or indirectly involve rocks and minerals. These challenges include the ever-decreasing supply of mineral resources, exploitation of mineral resources, environmental consequences of mining, etc. Sedimentary rocks are important for the students of Agronomy as the parent materials of most soils and raw materials for the manufacture of fertilizers and soil amendments.

Texture and structure are important diagnostic features of rocks. Rock structure is determined by the shape and size of mineral grains composing it. The mineral particles may be identified by a naked eye or under a microscope. The structure reflects the conditions of rock formation. Rock texture is determined by the way its component parts are distributed in space and their density.

Many intrusive igneous rocks are grainy in structure and coarse – textured or compact in texture, while many extrusive ones are porphyritic in texture with relatively large crystals (phenocrysts) surrounded by a "porridge-like" groundmass of small crystals.

The texture of sedimentary rocks is determined by the size, shape, and arrangement of the materials which form them. Some solidified-lava rocks are so porous that they are sponge like in appearance and feel (pumice). A glass-like structure is characteristic of obsidian (volcanic glass).

4.2. Igneous rocks

Igneous rocks are those that form from the cooling and crystallization of hot silicate liquids. The silicate liquids themselves are the product of partial or complete melting of the crust or mantle. Igneous rocks play an extremely important role in the evolution of the Earth. They constitute the bulk of the oceanic crust and comprise significant portions of the continental crust as well. Their formation is testimony to the high temperatures that exist to this day in the Earth's interior.

Magma is the term used to describe the liquid when it is within the Earth's crust or mantle. **Lava** is the term used to describe the liquid when it has reached the surface of the Earth. Sometimes magmatic materials are poured out upon the surface of the earth as when lava flows from a volcano. These molten materials are known as **eruptive, extrusive, or volcanic rocks**. Under other conditions, magmas do not come to the surface but may force their way or intrude into other rocks where they solidify. These intruding rock materials harden and form **intrusive** or **plutonic rocks**.

Igneous rocks may, among other things, be distinguished from sedimentary and metamorphic rocks by their texture, structure and complete lack of fossils.

Intrusive, or plutonic igneous rocks have been intruded or injected into the surrounding rocks. Intrusions of this type normally occur at great depth. Consequently intrusive bodies may be seen only after the ourlying rocks have been removed by erosion. Some of the more common intrusive bodies are discussed below.

Batholiths are the largest of igneous intrusions. They are irregularly shaped and may cover thousands of square miles. Batholiths extend great distances within the earth and become larger with depth.

Stocks. These are similar to batholiths, but cover an area of less than forth square miles.

Laccoliths are lenslike or mushroom-shaped intrusive bodies that have relatively flat under surfaces and arched or domed upper surfaces. They are intruded between the bedding planes and are thicker in the center and become thinner near their margins. Many of the domed mountains in the Crimea and Northern Caucasus (Beshtau, Mashuk and Zalizna) have been formed by laccoliths. To Russians they seem to be loaf-shaped.

Sills are tabular bodies of igneous rocks that spread out as essentially horizontal sheets between beds or layers of rocks.

Dikes. A dike is a tabular or wall-like mass of igneous rock that cuts across bedding planes when introduced into sedimentary rocks. Dikes commonly result from magma being injected into cracks and joints in the rocks, and range in size from a few feet to many miles in length. They are common in volcanic areas.

The fact that volcanoes exist and have erupted lava both in the geological past and at the present leads to the conclusion that temperatures within the Earth are sufficiently high for at least partial melting to occur. Geophysical evidence tells us that with the exception of the outer core, the interior is nowhere completely melted. Consequently, **magmas generated within the mantle must be the result of partial melting**.

Chemical and mineralogical composition of igneous rocks. Nearly all known chemical elements can be "discovered" in the igneous rocks but the most widespread of them are the so called petrogenic elements: O, Si, Al, Fe, Mg, Ca, K, Na, Ti, and H. Mean chemical composition of igneous rocks, according to Nickolds is given in a Table 10.

Table 10. Mean Chemical Composition of Igneous Rocks (after Nickolds), %

Rock	SiO_2	Al_2O_3	CaO	MgO	FeO	Fe_2O_3	Na_2O	TiO_2	K_2O	H_2O	MnO	P_2O_5
Andesite	54.20	17.17	7.92	4.36	5.49	3.48	3.67	1.31	1.11	0.86	0.15	0.28
Basalt	47.71	13.70	10.35	10.39	8.43	3.15	2.22	1.75	0.81	0.66	0.16	0.27
Gabbro	48.36	16.84	11.07	8.06	7.92	2.55	2.26	1.32	0.56	0.64	0.18	0.24
Granite	72.54	12.98	0.68	0.34	1.68	1.97	3.86	0.32	4.96	0.48	0.07	0.13
Diorite	51.86	16.40	8.40	6.12	6.97	2.73	3.36	1.50	1.33	0.80	0.18	0.35
Peridotite	43.54	3.99	3.46	34.02	9.84	2.51	0.56	0.81	0.25	0.76	0.21	0.05
Syenite	58.79	15.39	4.17	2.72	4.01	2.92	4.67	0.88	5.37	0.58	0.14	0.36

Depending on the silica content in the igneous rocks they are classified into 5 essential groups:

Ultra acid	>75% SiO_2
Acid	65-75% SiO_2
Medium (intermediate)	52-65% SiO_2
Basic	45-52% SiO_2
Ultrabasic	<45% SiO_2

Chemical composition of igneous rocks is entirely determined by their mineralogical composition. Thus acid rocks are composed prectominantly with quartz, orthoclase and mica, medium ones – with feldspars (including plagioclases) a little bit of hornblende, and biotite, whereas basic ones contain basic plagioclases and olivine. Ultrabasic rocks contain olivine, augite and metallic ore minerals. According to Bowen, minerals that crystallize (or melt) at high temperatures are rich in iron and magnesium and poor in silicon. Igneous rocks crystallizing these minerals will be **mafic** in composition (basalt and gabbro). The minerals crystallizing at the lowest temperatures are rich in silicon and aluminum, and rocks consisting of such minerals are called **felsic**. An important difference in magmas is their volatile content (chiefly the amount of H_2O). Mafic magmas are low in volatiles and felsic magmas contain significant amounts of volatiles. Mafic minerals and rocks are dark-colored, while sialic (acid) ones are light-colored.

Acid Intrusive Rocks

Granite is the most common acid intrusive rock. Magmas that do not rise to the surface but cool at depth are surrounded by heat-insulating rock. So they crystallize slowly, and consequently have the opportunity to grow large crystals.

Granite is a coarse-textured intrusive rock characterized by crystals which are all about the same size (from 1/16 inch to 1/2 inch or more) (Fig. 57).

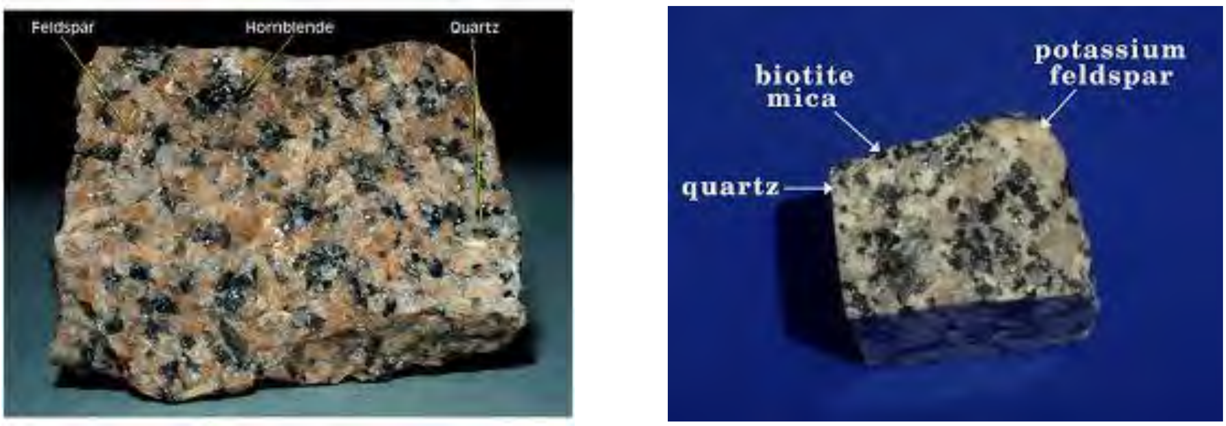


Fig. 57. Granite, a Coarse-Textured Intrusive Igneous Rock.

Granite always contains quartz and feldspar, usually intermixed with mica or hornblende. Aplite is a type of granite containing a great deal of orthoclase feldspar mixed with quartz.

Granites are typically light in color and may be white, gray, pink, red or yellowish brown. The individual mineral grains are easily distinguished, but are so well joined together that they form a hard, long-lasting rock. Because it is durable and can take a high polish, granite is especially useful as a building and monumental stone. Commercial granites are quarried in Zhitomir and Kirovograd regions.

Rapakiwi is a granite variety with a porphyritic structure, large orthoclase crystals are set in the frames of light-colored oligoclase, which is less resistant to weathering than the orthoclase so that the rock is transformed into a “rotten stone” (rapakiwi in Finnish).

Pegmatite (Arabian stone) is a granite variety occurring in pegmatitic veins. The rocks here cool at a very slow rate and form large crystals. The feldspar crystals are overgrown with those of quartz which creates the impression of the ancient Arabian (or Jewish) manuscripts with cuneiform writing. But sometimes the crystals may grow to several feet in length.

The high viscosity of sialic magma makes it difficult for such magmas to reach the surface. Consequently, rocks in this compositional range are most commonly found as intrusive varieties. Large granitic batholiths are typically found in tectonic regions undergoing mountain building and metamorphism.

Acid Extrusive Rocks

Such rocks may pour out of the craters of volcanoes or from great fissures or cracks in the earth's crust. This prevents slow crystal growth and results in a microcrystalline texture in which the crystals cannot be seen with the unaided (naked) eye. Some cool so rapidly that no crystallization occurs and this produces volcanic glass. The minerals rich in SiO_2 predominate in these rocks.

Liparite is an eruptive analogue of a granite. It possesses the same mineralogical composition but its structure is fine grainy, porphyritic or glassy. Phenocrystals are represented by an acid plagioclase, quartz and occasionally biotite and hornblende. Color of the rock is whitish grey, yellowish or pinkish. Specific gravity: 2.6. It is a good building and decorative stone.

Quartz porphyry. Some igneous rocks appear to have a mixed texture. This type of texture is called porphyritic and is characterized by relatively large crystals, called phenocrystals, surrounded by a groundmass (or background) of smaller crystals (Fig. 58).

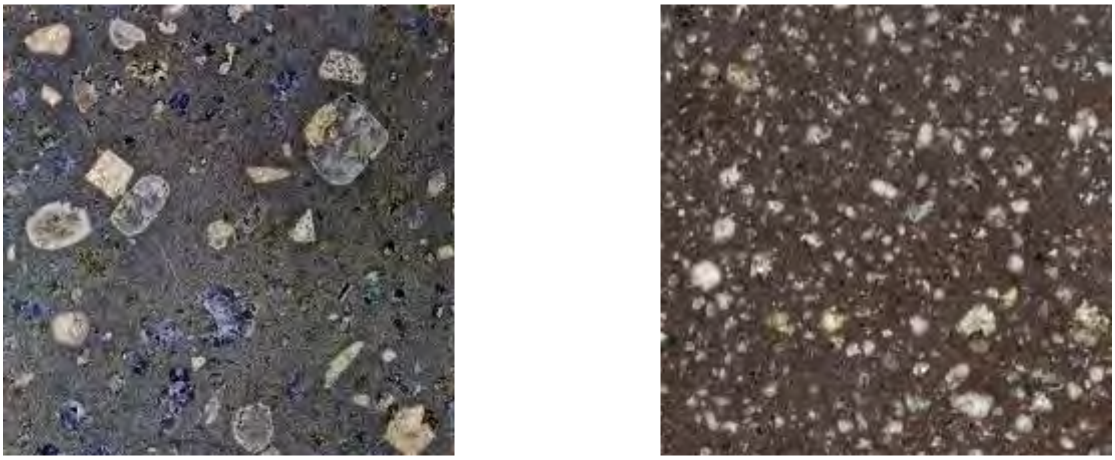


Fig. 58. Porphyry showing light-colored phenocrystals in the darker groundmass.

Porphyries are believed to represent two distinct phases of cooling and solidification. Due to its great strength (1300-1800 kg*cm^{-2}), this rock is employed in building and monumental construction.

Obsidian, known also as volcanic glass is a glassy extrusive rock which cooled so rapidly that it had no opportunity or time to grow crystals, even the small ones, at all and thus “froze” into a glass. It is a lustrous, glassy, black or reddish-brown igneous rock which exhibits a conchoidal fracture with sharp edges. Consequently this stone was commonly used by early man to make arrowheads, spear points, knives, and other implements. Even American Indians not so long ago used it for such purposes. Specific gravity: 2.2-2.3. Some beautiful varieties are employed as a decorative stone.

Pumice. Lava that solidified while steam and other gases were still bubbling out of it is called pumice. It is formed from a rapidly cooling volcanic froth and is characterized by the presence of large numbers of fine holes which give the rock a spongelike appearance. This rock is very light in weight and because many of the air spaces are sealed, pumice can float on water. Pumice is typically light in color and, though differing greatly in appearance, has the same chemical composition as obsidian a granite. It is used as an abrasive, in soap, cleansers, and some rubber erasers.

Medium (Intermediate) Intrusive Rocks

Syenite. Syenite resembles granite, but is less common in its occurrence and contains little or no quartz. If quartz is present the rock is referred to as a quartz-syenite. Consisting primarily of potash feldspars with some mica or hornblende, syenites are typically even-textured and the mineral crystals are usually small. Their deposits accompany those of granites or gabbro.

Diorite is a more common rock compared with the syenite. By its texture it is fine or medium grained. Plagioclases dominate in its mineralogical composition, but hornblende, biotite and even single grains of quartz may also occur. Quartz-diorites contain more quartz and look very much like some light-grey granites. Some deposits of iron and copper ores are connected with the diorites. Their strength is 1800-2400 kg*cm^{-2}.

Medium Extrusive Rocks

Trachites are composed of potash feldspars, plagioclases, biotite and hornblende. The structure is porphyritic, the phenocrysts including feldspars. Color is grey, yellowish or reddish. Strength is not so great (600-700 kg*cm^{-2}) due to porous structure. The rock is employed in construction.

Andesites are the extrusive analogues of the diorites, but their structure is porphyritic. Mineralogical composition: plagioclases, hornblende, less commonly augite and biotite mica. Color: brown or light brown. Andesites are common in the mountain regions. Alpine soils are formed on the products of their weathering in the Crimea and Carpathians. Their strength is great (1200-2400 kg*cm^{-2}). So the boulders of the rock are employed in construction and as road and street pavers. It is also an acid-proof material.

Basic Intrusive Rocks

Gabbro is a heavy, dark-colored igneous rock consisting of coarse grains of plagioclase feldspar and augite. Quartz is absent, and the mineral crystals are usually dark gray, dark green, or black. Sometimes biotite and very rarely pyrite may be encountered in the rock. In Zhitomir region of Ukraine there are large deposits of gabbro. It is a strong rock (2000-3000 kg*cm^{-2}) employed in monumental construction. Some beautiful gravestones are made of it.

Labradorite is a coarse-textured intrusive igneous rock quarried in Zhitomir region which may be very beautiful when polished, due to the play of colors. It is composed of a labrador and is a decorative stone.

Extrusive or Volcanic Basic Rocks

Basalt is one of the world’s most abundant extrusive rocks. Basalt rocks form the bulk of the oceanic crust. Basaltic lavas are erupted at mid-ocean ridges and subsequently move away from the ridge by sea floor spreading. In oceanic terraces, basaltic volcanism is dominant.

Basalts are typically dark gray, dark green, brown, or black in color and are normally quite heavy. They are fine-grained in texture and consist primarily of pyroxene, plagioclase feldspar, and, in some cases, olivine. Some basalts are characterized by a large number of open spaces or pores which mark the site of former gas bubbles. This porous rock, called scoria, is common in many hardened lava flows. With the passage of time these pores, or vesicles, may become filled with minerals such as quartz or calcite. Such mineral-filled vesicles, usually almond-shaped, are known as amygdales (amygdaloidal basalt). They often yield fine mineral crystals.

Basaltic rocks commonly display columnar jointing. This comes about as the rocks cool and shrink and split into vertical columns. Striking examples of these may be seen in many parts of the world including Ukraine (Rivne region, Ivanova dolyna) (scoria-вулканічний шлак). Basaltic rocks may be seen in the great lava flows of Hawaii; in some area of India and the north western United States there are large basalt flows covering about 300,000 square kilometers to depths of thousands of meters.

Because of their hardness, basalts are valuable as trap rock for road building and other constructive purposes. Large deposits of copper have been found in amygdaloidal deposits in northern Michigan. Some brown forest soils in the Carpathians were formed in the andesitic and basaltic products of weathering: the soils are of a characteristic reddish color.

4.3. Sedimentary rocks

All rocks, once exposed on the surface of the Earth are subjected to processes of weathering, erosion, transportation and ultimately deposition. Such processes produce **sediment** which may ultimately be transformed into **sedimentary rock**. To form sediments rock fragments are commonly picked up and transported by wind, water, and ice. Sediments are typically deposited in layers or beds called **strata**. When sediments become compacted and cemented together (a process known as **lithification**), they form sedimentary rocks. We repeat here that these rocks, represented by such common types as sand-stone, shale, loess and limestone, make up about 75% of the rocks exposed on the earth's surface.

Sedimentary rocks in this country are generally classified as **(1) clastic or detrital sediments; (2) chemical sediments and (3) biological sediments,** according to the source of the rock materials which form them.

4.3.1. Clastic Sedimentary Rocks

Clastic sediments are composed of rock fragments which have been derived from the decomposition or disintegration of igneous, sedimentary, or metamorphic rocks. Rocks formed from these worn-down rock particles are also called **detrital** or **fragmental** sedimentary rocks. Because the sediments which form these rocks are normally transported by mechanical means (wind, water, or ice), they have also been referred to as **mechanical sediments**. The erosion, transportation and deposition of material by moving water is an energy-controlled process that results in deposits of sediment that are typically well sorted in terms of the size of fragments. As a river slows down on entering to ocean, it will first deposit large materials, followed in turn by sand, silt and finally clay-sized particles. This process leads to the classification of the detrital sediments based on the size of the fragments that make up the sedimentary rocks. Table 11 shows the classification accepted in this country.

Table 11. Classification of Clastic Sedimentary Rocks

Particle size, mm	Coarse-fragment Rocks			
	Unconsolidated		Consolidated	
	Rounded	Angular	Rounded	Angular
>200	Bounded	Blocks	Boulder conglomerates	Block breccias
200-100	Large cobbles	Broken large stones	Large cobble conglomerates	Broken large stone breccias
100-50	Medium cobbles	Medium broken stones	Medium cobble conglomerates	Medium broken stone breccias
50-10	Pebbles	Fine broken stones	Pebble conglomerates	Fine broken stone breccias
10-5	Coarse gravel	Coarse angular gravel	Coarse gravel conglomerates	Coarse angular gravel breccias
5-2.5	Medium gravel	Medium angular gravel	Medium gravel conglomerates	Medium angular gravel breccias
2.5-1.0	Fine gravel	Fine angular gravel	Fine gravel conglomerates	Fine angular gravel breccias
Medium-fragment (Sandy) Rocks				
1-0.5	Coarse sand		Coarse sandstone	
0.5-0.25	Medium sand		Medium sandstone	
0.25-0.05	Fine sand		Fine sandstone	
Fine-fragment (Silty) Rocks				
0.05-0.001	Silts: loess and loess-like loams		Siltstones	

Very fine fragment rocks like clays are classified by some authors as clastic sedimentary rocks and by others – as chemical sediments. Cemented or hardened sediments of clay are called mudstone or shale. The process of converting unconsolidated sediment into consolidated sedimentary rock is known as **diagenesis**. This process involves compaction and squeezing as the sediments become buried. Diagenesis also involves cementing of the individual grains together as water is either evaporated or squeezed out of the primary pore spaces.

Some of the more common types of clastic sedimentary rocks are described below.

Coarse unconsolidated fragments like gravels and sands are composed of rock fragments and grains of minerals, among which quartz is dominant. Their water holding capacity is low and water penetrability – high. Soils formed in gravelly and stony rocks possess such properties. But they are well acrated and organic matter quickly decomposes in them. Some soils contain large rocks which obviously do not behave like soil although, if numerous, might affect the behavior of the soil in bulk. Where gravel and stones occupy enough of the soil's volume to influence soil physical properties significantly, their volume fraction and size range should be reported along with the specification of the finer soil material. The relatively inert gravel and stony fractions of soil particles are called the "soil skeleton", while the finer fractions (sand, silt and clay) are referred to as the "flesh" of the soil. Soils formed in the mountain regions, in alluvial and glacial deposits may be rich in the "skeleton" particles of soil texture.

The largest group of particles generally recognized as soil material is **sand**, which is defined as particles in diameter from 1 mm down to 0.05 mm (in Ukraine). The sand fraction is often further subdivided into subfractions such as coarse, medium, and fine sand. Sand grains usually consist of quartz, but may also be fragments of feldspar, mica, and occasionally heavy minerals such as zircon, tourmaline, and hornblende, though the latter are rather rare. In most cases, sand grains have more or less uniform dimensions and can be represented as spherical, though they are not necessarily smooth and may in fact have quite jagged surface (Fig. 59), which together with their hardness account for their abrasiveness.

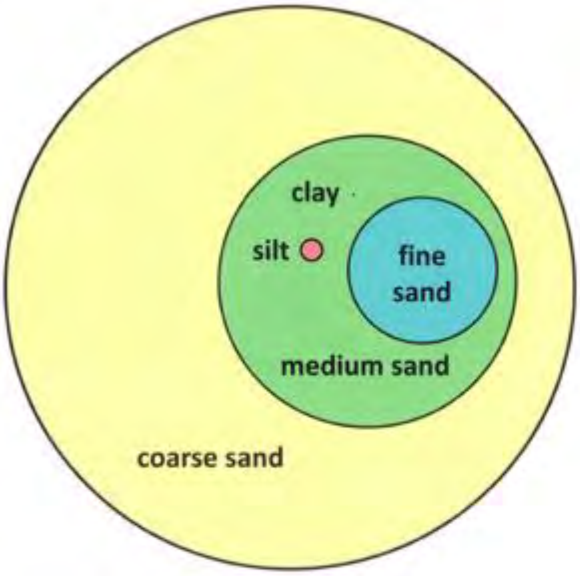


Fig. 59. A visual representation of the comparative sizes and shapes of sand, silt, and clay particles (source: www.ext.colostate.edu/mg/gardennotes/214.html).

As quartz particles have a hardness of 7, they will readily abrade steel (of hardness 5 ½), as can be commonly observed with tillage implements.

By their origin the sands may be alluvial (river valley sands), lacustrine (of the lake deposits), marine, fluvioglacial (glacial outwash sands) and aeolian (transported and deposited by the wind, especially in the deserts).

Their bulk density is within 1.4-1.7 and particle density: 2.6-2.7 (like that of quartz); total porosity: 30-50%; they have low capillary and high air-filled porosity; specific surface area is up to several square meters per gram (and usually much lower); infiltration capacity is high but water-holding capacity and the height of capillary rise are low. Such properties are imparted to the soils formed in sands.

Sandstone is composed essentially of cemented grains of sand. Sandstone has a granular texture and is the second most abundant of all sedimentary rock. In addition to quartz, sandstone may consist of sand-sized particles of calcite, gypsum, or various iron compounds. Arkose is a special type of sandstone which contains fragments of feldspar as well as quartz. Sandstone is used as an abrasive (for sandpaper) and as a building stone.

Conglomerate may be composed of rounded pebbles of many different sizes. It is essentially gravel which has been mixed with sand and held together by natural cement. The rock fragments forming a conglomerate may range from silt-sized particles to rocks the size of boulders. Conglomerates composed largely of angular pebbles are called **breccias**; those formed in glacial deposits (moraines) are called **tillites**.

Silt consists of particles intermediate in size between sand and clay. Mineralogically and physically, silt particles generally resemble sand particles, but since they are smaller and have a greater surface area per unit mass and are often coated with strongly adherent clay, they may exhibit, to a limited degree, some of the physicochemical attributed of clay.

The finer particles carried by the wind may accumulate to form deposits of silt known as **loess** [l æ s]. A yellowish, fine-grained, nonstratified material, loess is composed of small angular fragments of quartz and to a much lesser extent of a variety of other minerals. Ukrainian loesses usually contain over 50% of coarse silt particles (0.05-0.01 mm in diameter). Calcite content in some of them may be up to 30%. Of the other minerals, feldspars, micas, gypsum and iron oxides are the most common. The materials forming loess are derived from surface dust originating primarily in deserts, river flood plains, glacial outwash deposits, and deltas. Loess is quite cohesive and possesses the property of forming steep cliffs with vertical faces.

Loess is well known for its ability to form fine-textured, fertile soils. Such are Ukrainian Chernozems, grey forest and chestnut soils formed in loess or loess-like loams.

Loam is a sedimentary rock containing 40 to 70% of particles over 0.01 mm. Sand materials may be from 20 to 50% by weight. Loams may be of various origin, but very often they form on the slopes of mountains, hills, valleys and gullies as a result of a water flow and material transportation and deposition. Loess-like loams, unlike loesses, are stratified.

Shale. The most abundant of all sedimentary rocks, shale is formed from silt and clays which have hardened into a rock. Shale (Fig. 60) is characteristically fine-grained, thinly bedded, and split easily along bedding planes (the dividing planes which separate individual layers or beds of sedimentary rocks).

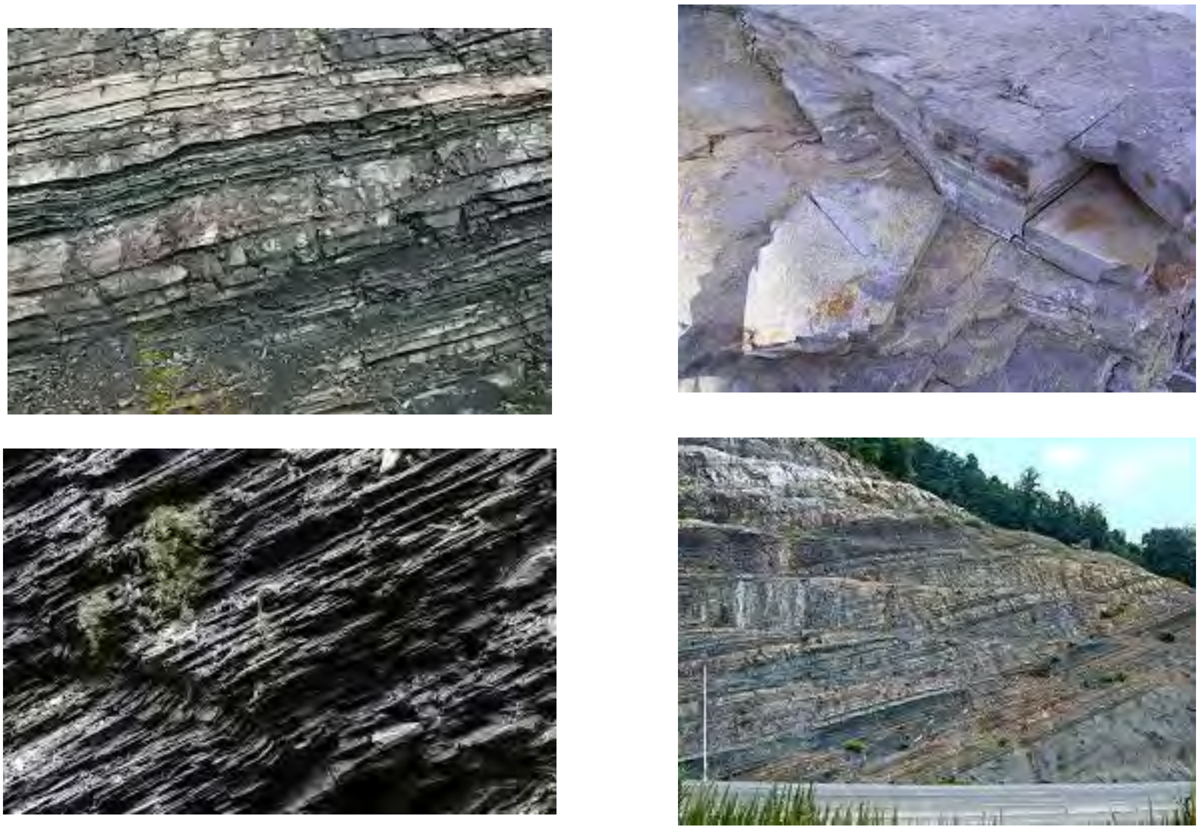


Fig. 60. Shale showing bedding planes.

Shale containing appreciable amounts of sand is called **arenaceous** shale; that containing large amounts of clay, or rather fine silt, is said to be **argillaceous**. **Carbonaceous** shale is typically black, and high in organic matter; shale which contains large amounts of lime is known as **calcareous** shale. Carbonaceous shale may yield petroleum or coal, and calcareous shale is used in the manufacture of Portland cement.

Clays. The clay with particles ranging from 2 mm downwards, contains much of a colloidal fraction. Clay particles are characteristically platelike or needlelike in shape and generally belong to a group of minerals called the aluminosilicates. In some cases, however, the clay fraction may include considerable concentrations of fine particles which do not belong to the aluminosilicates clay mineral category, e.g. iron and aluminum oxides and hydroxides. Americans differentiate between silicate and nonsilicate clays. Clays occupy about 50 % of the total area of sedimentary rocks.

Because of their great surface area per unit mass and resulting physicochemical activity, clays have a pronounced adsorbing (cation exchange and ion retention) capacity. Clay particles adsorb water and hydrate, swell upon wetting and shrink upon drying. A body of clay will typically exhibit plastic behavior and become sticky when moist and then crack and form cemented hard fragments when dessicated. All these and many other properties are imparted to the soils formed in clays.

All clay rocks are composed of particles less than 0.01 mm in diameter. They are “earthy”, adhesive and plastic. The essential clay minerals are Kaolinite, hydromicas and montmorillonite. Clays containing Halloysite, nontronite, allophanes, etc. are less common. Clays also contain silica, quartz, chalcedony, salts, sulfides, iron and manganese oxides and organic matter. Clays may be mono- and polymineral in composition. But the most common are kaolin and montmorillonite (smectites) clays.

Kaolins are monomineral clays, predominantly white in color with low plasticity, greasy to feel, sticky to tongue, with a characteristic odor of a cottage-house stove. Their formation is connected with the feldspar weathering. Ukraine is rich in kaolin deposits, formed as a cortex of weathering over the granitic shields. Kaolins are used in the production of aluminum and in the ceramic industries.

Montmorillonite clays are composed essentially of a montmorillonite (smectites) and the minerals accompanying it (nontronite, illite and beidellite). The clays are of white, light grey, yellowish and greenish color. They possess high adsorbing (and cation exchange) capacity. They may form as a result of weathering of basic, ultrabasic and carbonaceous rocks. Their plasticity, as well as the ability for swelling and shrinkage are very high. In Ukraine, the most capacious deposits of montmorillonitic clays are encountered in Cherkasy, Zakarpattya, Ternopil and Khmelnitsky regions. Smectites are common in Vertisols and soils of alluvial planes.

4.3.2. Chemical Sediments

Sediments which have precipitated from material dissolved in water are called chemical sediments. For example, rock salt may be precipitated from solution upon evaporation of sea water. Arid climate favors the process. Evaporites, limestones and dolomites are the most common among the chemical sediments. The first to precipitate are calcium sulfates (gypsum or anhydrite) then sodium chloride, potassium and magnesium salts come out of the solutions.

Rock salt is of crystal-grainy structure, monomineral but with usual impurities of gypsum or anhydrite (up to 30-40 %) sylvite, etc. It was formed in salt lakes and internal marine basins. In Ukraine its deposits are near the towns of Slovyansk and Artemivsk and near Kalush (Ivano-Frankivsk region). Rock salt (halite rock) is used mainly in the manufacture of hydrochloric acid, chlorine, soda carbonate and bicarbonate, sodium hydroxide and in organic synthesis.

Sylvite and other potassium salts are of importance as potassium fertilizers or a raw material for their production.

Gypsum and anhydrite are calcium sulfates which are used for the production of soil amendments. Very large deposits of gypsum are located at the Donetsky Highland. Extensive evaporite deposits occur in western Texas.

Limestone is composed primarily of one mineral: calcite ($CaCO_3$). There are many varieties of limestone, and some are formed by inorganic means, such as direct precipitation, while others are organic in origin. Travertine, which forms stalactites and stalagmites in caves, is a crystalline, usually banded variety of limestone. **Tufa**, a spongy, porous, inorganic limestone, is formed when calcite is deposited around springs and streams. **Dolomite** – known also as magnesium limestone, $[Ca,Mg(CO_3)_2]$ is formed when some of the calcium in limestone is replaced by magnesium. Ground limestone and dolomite are employed as aglimes to amend acid soils. Limestones or the products of their weathering may be parent materials of some soils especially the Rendzinas in the West of Ukraine and some ordinary Chernozems in the East of it.

Allithic rocks are composed of aluminum hydroxides. Bauxite (Al_2O_3*$2H_2O$) occurs in earthy, claylike masses, or in a **pisolithic** form, as rounded concretions in a claylike matrix. Hematite and other iron oxides may occur in it as impurities. The predominating colors are red and brown.

Ferrolythic rock is represented by limonite and ironous tufa. Some of such rocks may actually be used as iron ores. They form in the swamps of the Forest zone of Eurasia (taiga).

4.3.3. Biochemical or Organic Sediments

Biochemical (biological) sediments are formed from the remains and/or secretions of organisms.

Organic limestone – Most limestones are organic in origin and these rocks frequently contain the remains of the organisms responsible for their formation. Such limestones are composed of shells and coarse shell fragments. Reef limestones are composed of corals and other lime-secreting organisms, and chalk is a porous, fine-textured variety of limestone composed of the calcareous shells of microorganisms such as foraminifers.

Radiolarite is composed largely of the siliceous exoscelentons of tiny one-celled animals called rediolarians. **Diatomaceous earth** is formed primarily from siliceous remains of microscopic plants called diatoms.

Phosphate rocks contain 12 to 40 % P_2O_5. Ground rock phosphate is a source of phosphorus very slowly soluble in some acid and alkaline soils. The sedimentary deposits of phosphate rock were formed 10 to 15 million years ago and are buried under an overburden which may be very shallow, sometimes under 5 to 10 meters. Mining consists of removing the overburden and using a hydraulic gun to break up the ore to form a slurry. The slurry is pumped to a recovery plant, where rock phosphate pebbles are separated from sand and clay by a washing and screening process. The rock phosphate pebbles are ground to a powder, which is sometimes used directly as a fertilizer. Sometimes phosphorites form concretions around a fossil or some other nucleus. Concretions range from as little as one inch to several feet in length or diameter. Because concretions are usually harder than the enclosing rock, they are often left behind after the surrounding rock has been eroded away.

Caustobiolithes are the rocks capable to burn. They represent fossil fuels.

Coal is composed largely of carbonized plant remains. An important fuel for industry, coal is usually found in layers, associated with other sedimentary rocks. In its formation coal passes through several stages. **Peat**, composed of partially carbonized plant material, marks the first stage; **lignite or brown coal** is the second stage. Further stages may convert the lignite to bituminous or soft coal, and bituminous coal which has been metamorphosed will turn into **anthracite** or hard coal.

Histosols are organic soils formed in peats. Histosols contain organic soil materials to a significant depth. Fibrisols of Canada composed largely of relatively undecomposed fibric organic material were formed in peat deposits dominated by sphagnum mosses.

Fossils are the remains or evidence of ancient plants and animals that have been preserved in the earth’s crust. Fossils normally represent the preservable hard parts of some prehistoric organism that once lived in the area. Most fossils are found in sedimentary rocks.

Carbonaceous shale is typically black and high in organic matter. It may yield petroleum or coal. It is used as a fuel.

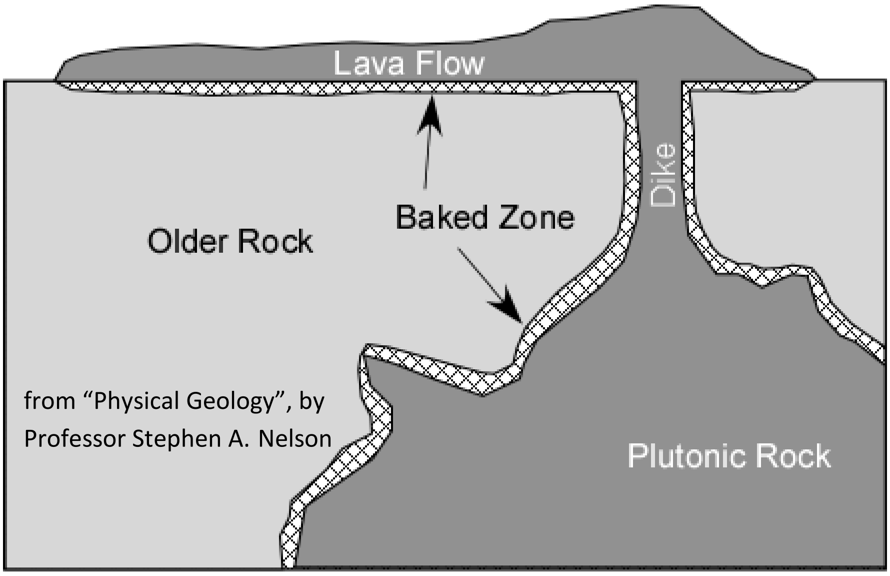
4.4. Metamorphic rocks

Metamorphic rocks are rocks (originally either igneous or sedimentary) that have been buried deep within the earth and subjected to high temperatures and pressures. The temperature becomes greater with depth. Its rise weakens the bonds between the atoms in crystal structures, increases the reactivity of water and other chemical compounds. Already at the depth of 10 km the pressure reaches 2600 atmospheres. These new physical conditions produce great changes in the solid rock and these changes are included under the term **metamorphism (Greek μετα, "change", and μωρφε, "form" or "shape")**.

During the process of metamorphism the original rock undergoes physical and chemical alterations which may greatly modify its texture, mineral, and chemical composition. Limestone may be metamorphosed into marble, and sandstone into quartzite.

Various classifications recognize several different **types of metamorphism**, but only contact metamorphism and dynamic, or kinetic, metamorphism will be considered here.

Contact Metamorphism is a local type of metamorphism that develops at the "contanc" of hot igneous intrusions and the surrounding rocks into which they intrude. The contact metamorphism is driven entirely by addition of heat into the surrounding rocks. Consequently this type of metamorphism is also referred to as **thermal metamorphism**. But the surrounding rocks undergo profound change. Limestone intruded by a hot magma may be altered for a distance of a few inches to as much as several miles from the igneous sedimentary contact. Since this environment is not associated with strong deformation, the metamorphic rocks in contact zones (also called **baked zones** or **contact aureoles**) usually do not exhibit foliation (Fig. 61).



from “Physical Geology”, by Professor Stephen A. Nelson

Fig. 61. Baked zone in country rock surrounding an igneous intrusion.

Physical change may be produced by contact metamorphism when the original minerals in the country rock are permeated by magmatic fluids which often bring about recrystallization. This process, which typically produces either new or larger mineral crystals, may greatly alter the texture of the rock. In addition, the magmatic fluids commonly introduce new elements and compounds which will modify the chemical composition of the original rock and result in the formation of new minerals.

Dynamic, or Kinetic Metamorphism occurs when rock layers undergo strong structural deformation during the formation of mountain ranges. The great pressures exerted as the rock layers are folded, fractured, and crumpled, generally produce widespread and complex metamorphic change. The major fold mountain belts are the Appalachians, Himalayas, Alps or Rocky Mountains. The metamorphism results from deep burial typically associated with the crystal thickening that results from thrust faulting and folding. The pressures may result in tearing or crushing of the minerals, obliteration of any indication of fossils or stratification, **realignment of mineral plains**, and increased hardness. In this type of environment, folding and differential stress are common; consequently **foliation** is a common feature of **regionally** metamorphosed rocks. This type of metamorphism is also called **regional metamorphism** as it takes place on a relatively large scale. So the effects of metamorphism produce the profound changes in the texture and chemical composition of the rock.

The single most distinctive feature of regional metamorphic rocks that separates them from both igneous and sedimentary rocks is the presence of a preferred orientation of the individual minerals that comprise the rock. Thus micas will tend to grow with their flat sheets perpendicular to the maximum compressional stress. There are three basic types of foliation: slaty, schistose and gneissic ['naisik].

The metamorphic equivalents of some common igneous and sedimentary rocks are given below:

ORIGINAL ROCK	METAMORPHIC ROCK
Sedimentary	
Sandstone	Quartzite
Shale	Slate, phyllite, schist
Limestone	Marble
Bituminous coal	Anthracite coal, graphite
Igneous	
Granitic textured igneous rocks	Gneiss
Compact textured igneous rocks	Schist

Slate. A metamorphosed shale, slate is characterized by a very fine texture in which mineral crystals cannot be detected with the naked eye. It does not show banding (Fig. 62). It splits readily into thin even slabs.

Fig. 62. Slate, a foliated metamorphic rock.

Slate occurs in a variety of colors, but is usually gray, black, green, and red. Its characteristic slaty cleavage (not to be confused with mineral cleavage) makes it specially useful for roofing, blackboards, and sidewalks.

Schist is a medium to coarse-grained foliated metamorphic rock formed under greater pressures than those which form slate. It consists principally of micaceous minerals in a nearly parallel arrangement called **schistosity**. Schists usually split readily along these schistose laminations or folia, which are usually bent and crumpled. Commonly derived from slate, schists may also be formed from fine-grained igneous rocks. They are named from fine-grained igneous rocks. They are named according to the predominant mineral, such as **mica schists, chlorite schists,** etc.

Phyllite. Derived from the Greek word phyllon (a leaf), phyllites are more fine-grained than schists but coarser than slate. On freshly broken surfaces they have a characteristic silky luster or sheen due to the presence of fine grains of mica. Most have been formed from shales which have been subjected to pressures greater than those required to produce slate, but not of sufficient intensity to produce schists. The only deposit of pyrophylite slates in the world occupying an area of some thousands of square kilometers is located in Zhitomir region near the village of Zbranka.

Gneiss (pronounced “nice”) is a very highly metamorphosed coarse-grained banded rock. This rock is characterized by alternating bands of darker minerals such as chlorite, biotite mica or graphite (Fig. 63).

Fig. 63. Gneiss, a banded metamorphic rock.

The bands are typically folded and contorted, and although some gneisses resemble schists, they do not split nearly as easily. Banding may be an indication of stratification if the original bedded sedimentary rock, or caused by the alteration of coarse-grained igneous rocks containing light – and dark – colored minerals. In general, gneiss have undergone a greater degree of metamorphism than have schistose rocks and are commonly formed as a result of intense regional metamorphism.

Nonfoliated Metamorphic Rocks are typically massive or granular in texture and do not exhibit foliation. Although some nonfoliated rocks resemble certain igneous rocks, they can be differentiated from them on the basis of mineral composition.

Quartzite is formed from metamorphosed quartz sandstone. One of the most resistant of all rocks, quartzite is composed of a crystalline mass of tightly cemented sand grains. When formed from pure quartz sand, quartzite is white; however, the presence of impurities may stain the rock red, yellow, or brown. Pink-colored quartzites from Zhitomir region of Ukraine are used as decorative stones. A well known Lenin mausoleum in Moscow had been faced with them.

Marble is a relatively coarse-grained, crystalline, calcareous rock, a metamorphosed limestone or dolomite. It is formed by recrystallization, and any evidence of fossils or stratification, to say nothing of the ripple marks in limestone, is usually destroyed during the process of alteration. White when pure, the presence of impurities may impart a wide range of colors to marble. In ancient Greece and Rome and in a more recent periods, some beautiful sculptures were made of marble, including the Venus of Taurida, the Statue of Athena, Heracles the Lion-Slayer and the later works of Michelangelo Buonarrotti (1475-1564). Some less “pedigreed” varieties of marble like the “sugar-like” marbles of Zhitomir region are employed as decorative stones.

Questions and Assignments

1. What are rocks and how are they classified?
2. What is the difference between extrusive and intrusive rocks?
3. Name the principles of igneous rock classification.
4. Describe granite (gabbro, peridotite and syenite).
5. Which is one of the world’s most abundant extrusive rocks? Describe it.
6. What volcanic rocks do you know? Describe pumice and obsidian.
7. How are sedimentary rocks classified?
8. Clastic sedimentary rocks and their classification.

9. Inorganic chemical sediments and their role in Agronomy.
10. Biochemical (organic) sediments and their role in Agronomy.
11. Phosphorites (phosphate rock) and their use in the manufacture of fertilizers.
12. What is contact metamorphism and baked zone of the altered country rock?
13. What foliated and nonfoliated metamorphic rocks can you name?
14. Describe gneiss (marble).
15. Which rocks and why are the most common as parent materials of soils?
16. How are peat and coal related to each other?

Chapter 5

Elements of Geochronology

Many attempts have been made to measure the age of the earth. Some of these attempts are based on relative amounts of radioactive elements within the rocks. Scientists assign an age to the rock layers containing these elements. They attempt to judge ages for other layers by their relationships with the dated layers.

The current estimated age of the earth is 4.5 billion years. This is not a guess, but based on natural explanations. Geologists study layered rocks for clues to their age. They also study fossils, the remains of plants and animals, found in the rocks. Fossils usually are found only in sedimentary rocks. Much of our thinking about geologic history is based on what we learn from the sedimentary rocks.

5.1. Dating Geologic Time

5.1.1. Measuring Earth's Age

Most sedimentary rocks are laid down in horizontal layers. The oldest beds are at the bottom. As age goes on, younger beds are laid down on the older ones. This idea is called the law of superposition. By this law the geologists determine the relative age of the rock by its position (its layer position) among the surrounding layers. To know the relative age is not enough. We need more information than just the rocks' position to tell how long ago they were formed.

If continents had stayed in place, their history might be easier to unravel. But when continents bump into one another or into ocean plates, rock layers are folded and faulted. Then sedimentary layers may no longer be horizontal, and relationships among them may be very complex. Even where sedimentary layers are uplifted, faulted, and folded, the order among them may be determined. We do this by following by layers to a place where they are still horizontal. Sometimes the presence of fossils helps to determine which layers are the oldest. Relationships among igneous rocks are more difficult to determine than among layered rocks. The first step is to see how igneous bodies are related to sedimentary rocks. If an igneous body cuts across a series of layered rocks. It is clear that the layered rocks were there first. Where layered rocks have been pushed up by igneous bodies, we know the layered rocks are older than the igneous rocks. Volcanic ash and lava flows are igneous rocks that may be useful in dating. Both of these deposits may be widespread, and sometimes they contain radioactive elements. Unconformities are breaks in the rock record. They are useful clues to the relative ages of layered rock. Unconformities are erosional surfaces separating series of deposits. From unconformities it is evident that lower layers were exposed above sea level long enough to be weathered and eroded. Once again the beds were covered by the sea and a new period of deposition followed.

Sometimes folded rocks are present beneath an unconformity. These folded layers may be covered by horizontal beds, which were not present when folding occurred. This series of events includes deposition of the lower unit, followed by folding, uplift and erosion. Again the area was covered by the sea. Deposition of younger horizontal beds followed. Unconformities also may separate two sets of horizontal beds. In such areas, the land has been under water, uplifted, and eroded. Later the region has been submerged again, and younger layers deposited. If enough time goes by during erosion, fossils in the layers beneath the unconformity may be much different from the fossils above the erosion surface. Major changes in the fossil content of the rock layers are used to separate geologic time into units.

Relative dates place events in historical order by comparing one with another. We cannot tell how many years have passed between events. Relative rock ages are based on relationships among the various beds. Lava flows between sedimentary beds tell us that the beds below are older than the flow. The beds above the flow must be younger than the lava.

5.1.2. Radioactive Dating

Measurement of time must be based on an accurate scale. One source of information used to determine the actual (absolute) age of rocks is radioactive dating. Radioactive materials such as Uranium-238 or carbon-14 has a specific half-life (the amount of time it takes for half the material to change to something else – see the following Table). These radioactive materials are found in rocks in small amounts. We can use this half-life to measure the age of material in a rock by seeing how much of the material is left after a given amount of time. The geologists compare the amount of parent material and daughter product to tell how much time has passed since a rock was formed.

Table 12. Frequently Used Radioactive Isotopes

Parent Material	Daughter Material	½ Life
uranium-238	lead-206	4.5 billion years
uranium-235	lead-207	7.3 mln yrs
thorium-232	lead-208	14.1 billion yrs
rubidium-87	stroutium-87	47 billion years
potassium-40	argon-40	1.3 billion yrs
carbon-14	nitrogen-14	5730 yrs

Radioactive parent material is said to decay when its stable nuclei break apart. The material remaining is called the daughter product. The working equation of radioactive dating looks like that:

$$t = \frac{T}{0.693} ln \frac{M_0}{Mt}, \qquad (3)$$

Where t = age in years; T = half-life period in years; M_0 = the amount of radioactive isotope at zero time; Mt = the amount of the same radioactive isotope at time t. Radioactive activities may be used instead of amounts (masses) of radioactive isotopes.

If uranium-238 is present in a rock, its age can be determined using the following formula:

$$Age = \frac{amount\ of\ lead-206}{amount\ of\ uranium-238} \times 7.6 \times 10^9 \qquad (4)$$

One gram of uranium-238 produces $\frac{1}{7.6 \times 10^9}$ g of lead-206 per year.

In our country to determine the absolute age of rocks the following methods are employed:

- Lead method for the rocks containing uranium and thorium minerals.
- Helium method, which allows to determine the age of minerals by the amount of helium formed together with the isotopes of lead during the radioactive decay.
- Argonmethod, which allows to find out the age of potassium-containing minerals and rocks. Potassium-40 is transformed into argon-40. Argon, unlike helium, can be retained in the minerals. Potassium is a component of nearly all rocks.
- Rubidium-strontium method. As is clear from the Table, given above, rubidium-87 is transformed into strontium-87 with a half-life period about 50 billion years. To determine the age of geologic formations, micas and other silicates are taken for measurements.
- Carbon method is used to determine the age of archeological wooden objects, plant residues, “old” trees, and even soil organic matter (SOM).

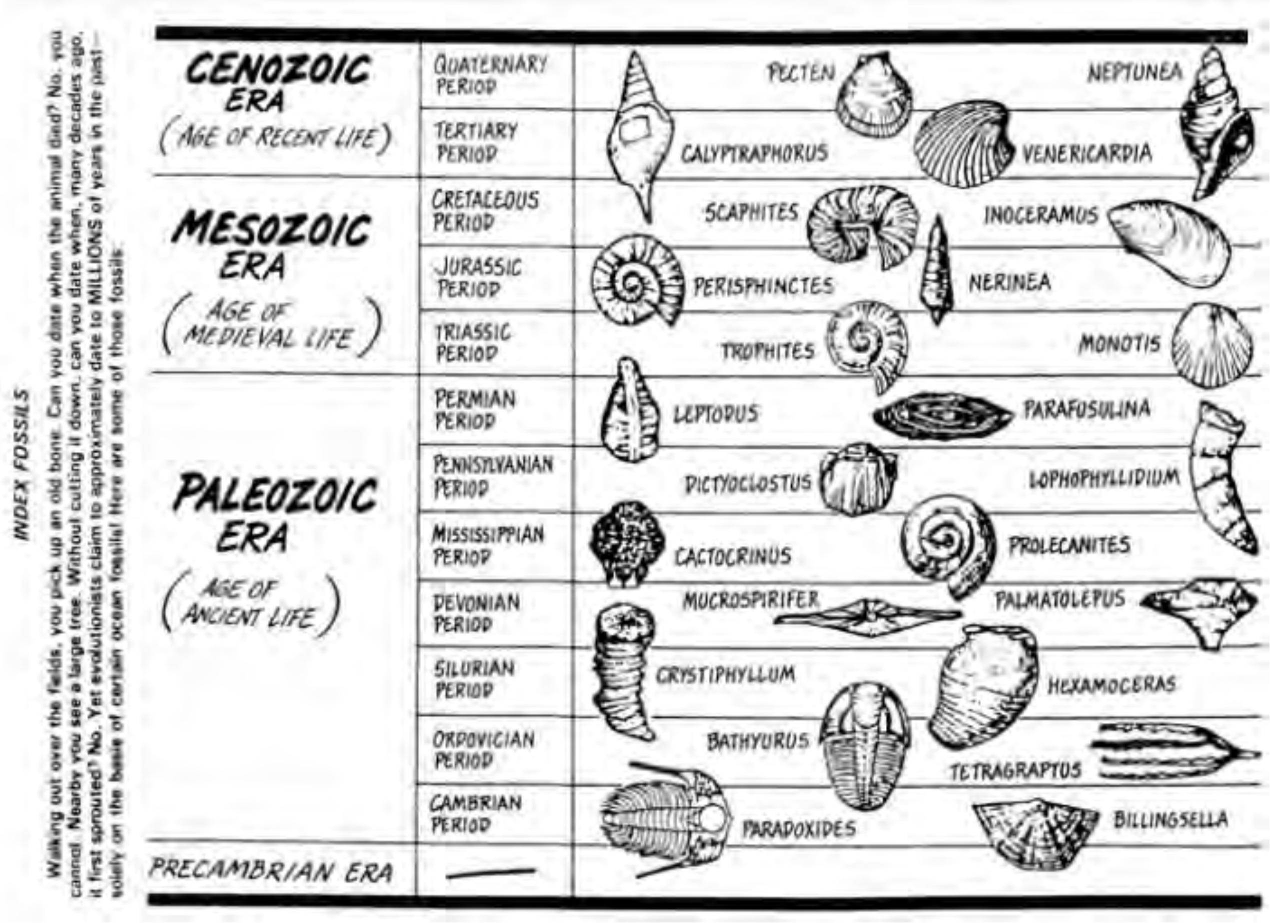
5.1.3. Fossil and Sedimentation Clocks

Once again about our country. The relative age of geologic rocks is determined by the use of the following methods:

Stratigraphic method uses the law of super position: younger beds are laid down on the older ones. Many complicated circumstances are possible here.

Lithologic and petrologic method takes count of the sets of layers and their comparison with the sets of layers in other regions, the relative age of which is well known.

Paleonthologic method studies the fossils. One form of fossilized animals and plants is evolutionarily changed into other forms. Their sequence means a certain evolutionary progress of living forms. There are certain leading forms of fossils. Sometimes entire animals are preserved, as happened in Siberia where wooly mammoths were buried and frozen in the ice. Other types of presentation are carbon impressions. Leaves and soft-bodies animals occasionally are buried in soft mud. Foot prints are another type of fossil record. Tracks of animals may be left in soft mud. Teeth seem to be preserved better than other parts of the body. Mineral matter may seep into bone openings, into shells, or into wood openings. Minerals matter then may harden into time units based on changes in the fossil record. Division of time are based on the appearance and disappearance of certain fossils. Fossils, which divide geologic time into small units, are called guide or index fossils (керівні викопні форми).



INDEX FOSSILS

Walking out over the fields, you pick up an old bone. Can you date when the animal died? No, you cannot. Nearby you see a large tree. Without cutting it down, can you date when, many decades ago, it first sprouted? No. Yet evolutionists claim to approximately date to MILLIONS of years in the past—solely on the basis of certain ocean fossils! Here are some of those fossils:

Fig. 64. Index fossils.

Beside fossil clocks, the geologists are using sedimentation clocks. This method uses the present rate of deposition of sediments to estimate how long it took to deposit the thickness of sedimentary rocks found on Earth. This method has many weaknesses, but it comes surprisingly close to some radioactive dates.

Geologists work out the history of events from layered rocks in local areas. Then they attempt to match this history with layered rocks of nearly areas. These known layers then are matched with layers of more distant areas. From this work, geologists have arranged all known rock layers in the order in which they are deposited. This sequence of beds is known as geologic column. The law of superposition is the basis for the geologic column. Geologists also use fossils to be sure that they have interpreted the layers in the right sequence.

Facies [FAYsheez] fossils are remains of generations of organisms that lived for relatively long periods but only in certain environments. Facies fossils occupy many different environments. They live for long periods of time if the environment remains the same. They die when the environment changes.

Facies fossils are not very helpful in dating time. However, they help up understand past environments. Dinosaur bones indicate land environments. Bones of swimming reptiles tell us the location of ancient seas.

5.2. Changing Patterns of Life

The theory of evolution was proposed by Charles Darwin (1809-1882) and Alfred Wallace (1823-1913). Some changes in organisms are brought about by mutations. These changes in the gene material cause an offspring to differ from its parents. Changes due to mutations are passed on then from the parent to its offspring. Mutations are sometimes harmful and the species may die out. Some mutations are useful and help the species to survive changes in the environment.

A species is a group of very similar organisms. Individuals of the groups are never identical. Small individual differences, however, may lead gradually to the development of new species. The horse changed from a four-toed animal, through three toes with the middle one touching the ground, to its present one toe or hoof.

Another important influence on the development of different species is the effect of isolation. On a small scale, Darwin studied isolation on the Galapagos Islands.

Continental drift brought about the separation of continents during the last 65 million years. Unique groups of animals developed in South America, Africa, and Australia due to their isolation.

In addition to biological reasons for changes in the history of life, there may be some geological causes, especially for extinctions. With a retreat of the sea and an increase in land area, many new environments are available. Land plants and animals would probably increase in number and in varieties.

5.3. The Geologic Time Scale

Earth has been in existence about 4.6 billions of years. However, we have not found rocks this old. The first billion or so years of Earth history have been lost from the record due to burial by younger rocks or erosion. Earth's oldest rocks, found in Greenland and Africa, are only about 3.5 billion years old. Some place within the ancient rocks, the first records of life on Earth probably are buried. Our earliest life records consist of microfossils with bacteria like primitive forms found in African rocks 3.2 and 3.4 billion years old. Since this life capable of photosynthesis had developed, we know that sometime during the earth's first billion years life began. This life already was changing Earth's atmosphere from one without oxygen to one rich in oxygen. With records of life development, it becomes possible to divide geologic time into units which we call the geologic time scale.

Table 13. Geologic Time Scale.

Era	Period	Epoch	Life	Age Estimate (absolute), mln. years
Cenozoic KZ	Quaternary Q Pleistocene	Recent (Holocene) Q_4 Late quaternary Q_3 Middle quaternary Q_2 Early quaternary Q_1	The appearance of man. Development of present-day flora and fauna. Golden age of mammals, bony fishes and insects.	1.5 … 2
	Neogene N	Pliocene N_2 Miocene N_1	Golden age of angiosperms. The appearance and development of mammals close to the present-day forms. Development of anthropoid primates.	26±1
	Paleogene P	Oligocene P_3 Eocene P_2 Paleocene P_1	Dying out of Mesozoic fauna and flora. Golden age of primitive mammals.	67±3

Era	Period	Epoch	Life	Age Estimate (absolute), mln. years
Mesozoic MZ	Cretaceous K	Late Cretaceous K_2 Early Cretaceous K_1	The appearance of angiosperms, and the bloom of gymnosperms. Golden age for large reptiles. Development of mammals and birds.	137±5
	Jurassic I	Late Jurassic I_3 Middle Jurassic I_2 Early Jurassic I_1	Golden age of gymnosperms (palmsand coniferous plants); gigantic lizards reign supreme. Flying lizards and birds appear.	195±5
	Triassic T	Late Triassic T_3, Middle Triassic T_2, Early Triassic T_1	Extinction of tree-ferns. Dominance of gymnosperms gradually declines. Development of conifers. Bony fishes. Dinosaurs.	240±10
Paleozoic PZ	Permian P	Late Permian P_2, Early Permian P_1	Gradual decrease of ferns and horsetails. Dominant plant forms are gymnosperms appearance of the first conifers. Development of insects, reptiles. Therapsida = the ancestor of mammals.	285±10
	Carboniferous C	Late Carboniferous C_3 Middle Carboniferous C_2 Early Carboniferous C_1	Enormous club-mosses, tree-forms, horsetails. Gradual expansion of gymnosperms. Great forests of tree-ferns and gymnosperms. Appearance of insects	340 … 360

Era	Period	Epoch	Life	Age Estimate (absolute), mln. years
			resembling dragonflies and cicadas. Ascendancy of achinoderms (see lilies). Considerable distribution of ancient sharks and amphibians appearance of forms between amphibians and reptiles. First reptiles appear.	
	Devonian D	Late Devon D_3 Middle Devon D_2 Early Devon D_1	Real forests cover the land. Wide flourishing of tree-like horsetails; first gymnosperms appear. Marine arachnids dominate. Further development of fishes. First known imprints of a five-toed ancestor of modern amphibians (Stegocephalus). Extinction of trilobites.	410 ± 10
	Silurian S	Late Silurian S_2 Early Silurian S_1	First authentic trace of land vegetation (tiny spore plants resembling modern ferns). Algae dominate. First land arthropods (closely resembling modern scorpions and wingless insects). Development of fish.	440 ± 15
	Ordovician O	Late Ordovic O_3 Middle Ordovic O_2 Early Ordovic O_1	Numerous species of marine algae, probably appearance of first	500 ± 20

Era	Period	Epoch	Life	Age Estimate (absolute), mln. years
			land plants. Dominant form-trilobites (fossil crustaceans). Numerous forms of corals, trilobites, molluscs (cephalopods), first chordates.	
	Cambrian E	Late Cambrian E_3 Middle Cambrian E_2 Early Cambrian E_1	Bacteria, blue-green algae, fungi. Existence of all ancient representatives of modern phyla except chordates.	570±30
Proterozoic PR	Late Proterozoic PR_3 (Rypheic)	Vendic V Late Rypheic R_3 Middle Rypheic R_2 Early Rypheic R_1	Various bacteria, blue-green algae. Unicellular organisms. At the end of the era the first appear-sponges, jellyfish, worms, mollusks, etc.	680±50 1050±50 1400±50 1600±50
	Middle Proterozoic PR_2	Not divisible		1900±50
	Early Proterozoic PR_1			2600±50
Archean AR	Not divisible	Not divisible	Fossil remains are absent for the greater part of the era. Hypothetical existence of primordial forms of life.	2600

We are now able to speak with great precision not only of geological periods, but also of the evolution of organic forms and of separate groups of these forms.

All geologic time up to 600 million years before the present is called Precambrian. This long time span has so few fossils that the only subdivisions made are based on radioactive dates. After 600 million years life changed relatively fast. This second stretch of time is subdivided into three eras. They are called the

Paleozoic, ancient life; the Mesozoic, middle life; and the Cenozoic, or recent life. Each era is then further divided into periods and periods into epochs.

5.4. Geologic Time Characterization

5.4.1. Precambrian Time

This time lasted from the beginning of the earth to about 600 million years before the present. Earth's primitive atmosphere must have had the six elements which are among the most abundant in the solar system: hydrogen, carbon, nitrogen, phosphorus, oxygen, and sulfur. These elements include the principle constituents, or necessary substances, for life. The bacterialike life found in these early rocks used carbon dioxide from the primitive atmosphere and sunlight to produce food. From an atmosphere with no free oxygen, the air gradually became oxygen rich. Then the abundance of oxygen made possible many changes in the development of new forms of life.

An ozone layer developed that protects life from too much ultraviolet radiation. By about 1.4 billion years ago, organisms were large, probably with cell nuclei. Organisms similar to blue-green algae were abundant. Oxygen released by the life in the rocks combined with iron dissolved in the ocean. The iron oxide which formed was then deposited from the sea to form Earth's greatest reserves of iron ore. Precambrian rocks form the nucleus of each continent.

We can also recognize a number of mountain building events. These early mountains were worn down, only to be uplifted again. Most of the exposed Precambrian rocks are igneous or metamorphic. Very few are sedimentary.

5.4.2. The Paleozoic Era

The Paleozoic Era began about 600 million years ago. Many fossils were left in most sedimentary layers. Early Paleozoic animals were marine invertebrates, animals without backbones. They probably lived near shore in shallow water. Their body coverings were like those of modern shrimp and soft-shelled crabs. The most important animals were trilobites and brachiopods. All the basic animal groups began to evolve. During the early Paleozoic, the continents began to break up and drift apart. The one continent in the Precambrian became possibly three continents in the Northern hemisphere and one continent in the Southern hemisphere. Shallow seas invaded the continents.

By middle Paleozoic time, marine life forms had developed shells of calcite. These shells were presented in great numbers. Great limestone reefs were built by corals and others marine organisms.

Mountain building began as European and North American plates joined. During the uplift, major changes occurred in both plant and animal kingdoms. Trees developed from land plants that had appeared earlier. Giant ferns and march plant were forerunners of today's vegetation. As land plants became available for food, animals moved to the land. Fish became the dominant occupants of the sea. One kind of fish developed a lung that allowed it to survive out of water. The fish also had fins that would support its weight. It became the ancestor of amphibians. They cannot live far from water. Their eggs, which have no protective covering, must be laid in water. Modern amphibians include frogs and toads.

During late Paleozoic, amphibians became numerous and dominated life on land. One variety of amphibians developed an egg with a hard covering. This type of animal is called a reptile. Reptiles do not have to return to water to produce their young. Reptiles also developed hard scales to prevent loss of their body fluids. These two developments allowed reptiles to adapt to a variety of environments. Like fish and amphibians, reptiles are cold-blooded animals. They must spend long hours in the sun to maintain their body temperatures. During late Paleozoic time inland seas were cut off from the ocean. About 225 million years ago, continental plates drifted together once more. The Earoamerican continent once more joined Gondwanaland, the southern continent. Mountains rose during this period. The Appalachian Mountains formed between the Earoamerican plate and the Gondwana plate, and he Hercynia [hɔr'sainiəu] Mountain belt rose in Europe. The Ural Mountains formed as the European and the Asian plates met. Interior deserts spread over much of the United States and parts of Europe during this time. Great salt beds formed in what is now Germany and West Texas.

As these events affected the continents, mass extinctions of both land and sea animals occurred. 75 percent of the amphibian families and about 80 percent of the reptile families disappeared. This time is often referred to as the "great dying".

5.4.3. The Mesozoic Era

In early Mesozoic time, about 190 million years ago, Laurasia and Gondwana again became separate continents. Between these continents, a sea formed. In the late Mesozoic, seas spread inland and covered many continental areas of Europe, North and South America. Plants of the Mesozoic had to adapt to an unstable environment in place of warm, humid climate of the Paleozoic. Angiosperms, the flowering plants we know today, developed during this time. They had seeds that could survive because of their hard, outer covering. The seeds could live through hot, cold, wet, or dry seasons, and sometimes for many years. Marine animals of this time were dominated by the ammonites. These invertebrates have distinctive markings on their shells which make them excellent index fossils.

Dinosaurs are the most unusual and interesting life form of the Mesozoic. A current theory suggests that they were warm-blooded ancestors of the birds. The largest one weighed as much as 17,000 kilograms. Tyrannosaurus was the largest carnivorous, or meat eating, land animals it stood about 6 meters tall and was 15 meters long. One flying reptile, discovered in West Texas, had a wingspan of about 15.5 meters. Birds and mammals also appeared during the late Mesozoic.

By 70 million years ago, continents were once more breaking up. Eurasia survived as a unit, but North and South America, Antarctica (including Australia), Africa, and India became separate plates. The plates drifted northward, eventually reaching their present positions. Seas drained off the lands, and mountain building gradually lessened. Dinosaurs and many other life forms died off or became less dominant.

5.4.4. The Cenozoic Era

It began when the ammonites and dinosaurs disappeared. As continents were uplifted, many new life forms appeared on the different continents. Early in the Cenozoic Era, there were great flows of basalt forming the Columbia River Plateau. The Rocky Mountains were uplifted as were the Coast Ranges in North America and the Andes Mountains in South America. Greenland, part of the American plate, began to drift away from Norway opening a water connection between the Arctic Sea and the Atlantic Ocean.

The Alps Mountains of Europe were uplifted during the Cenozoic Era when Africa drifted into Eurasia. The collision of the Indian, Chinese, and Arabic plates with the eastern edge of the Eurasian plate uplifted the great Himalayan Mountains. Australia and South America were separated from Antarctica and drifted into their present positions. Separation seems to have changed the development among the mammals found in Australia and South America.

Marsupials [ma’sjupi:əlz], like the kangaroo, became the dominant animal. For a while, marsupials (сумчасті) were common in South America. But in the late Cenozoic Era, the Isthmus of Panama formed a land bridge between North and South America. Animals from these continents could migrate from one place to the other. The armadillo [ama’di:loh] (броненосець), however, moved north and now inhabits much of the southwest United States.

The dominant marine invertebrates in the Cenozoic Era were members of the snail, starfish, and clam groups. Fish still dominated the sea. Birds dominated the air.

Mammals dominated the land. Birds and mammals are both warm-blooded animals. Thus, they could exist in some areas where reptiles could not live. Early protection of their young helped mammals survive in many, different environments. The ancestors of some mammals such as the whale returned to the sea. Others, like the bat, developed the ability to fly.

A major event of the late Cenozoic Era was the invasion of the northern hemisphere by glaciers. Glaciers reached south to the present location of the Ohio River. In Europe, ice covered much of what is now Scandinavia, Scotland, Germany, and Russia (Ukraine included). The glaciers repeatedly advanced, melted, and then advanced again. This ice age may have lasted about two million years.

During the ice age, humans left their marks on the land. Some of the earliest records of humans are stone tools. We really have very few records of early people. Their history is based on small amounts of evidence. A few nearly complete skeletons have been found. But more often, only a few skulls, thighbones records of early human groups are found in the caves of southern France and Spain. Paintings of animals can be seen on the walls of the caves. These caves are dry and the paintings have been preserved.

5.5. Development of the Earth's Crust in Quaternary Period

Quaternary period began 1.5-2 mln years ago and lasts to this day. It is divided into four epochs: early, middle, and late pleistocene, and holocene. A man appeared in this period, so it is also called anthropogenic. In the later epochs of this period the tools were found as well as creations of art – not only cave paintings, but also the figurines of animals and human beings modeled from stones and bones. Glaciations are the most important events of the quaternary period. The last glacier left the territory of Ukraine some 10 thousand years ago. The European part of Eurasia had four major glaciations: likhvin, dnieper, moscov and valday. During the most potent (dnieper) glaciations, the thickness of ice reached 2 kms, whereas the area of glaciations was about 5.5 mln km^2. During the epochs of glaciations, the level of water in the rivers, lakes and seas decreased which lowered the basics of erosion, caused regression of sea level and increased denudation. Marine and river terraces were formed and stream systems changed their outlines.

During glacial melting, numerous water streams appeared which transported clastic material. River levels increased as well as geologic work of rives. Sea level increased as well as geologic work of rivers. Sea level increased too. Alternating glacial and interglacial periods, transgressions and regressions, lifting and lowering of erosion bases – all these formed new landforms in series of sedimentary rocks which were called athropogenic deposits. Quaternary period geochronology may be expressed as follows:

(1) the appearance of man: 1.5 mln yrs;
(2) the beginning of glacial epochs: 1 mln yrs;
(3) the end of glacial epoch: 10-12 thous. Yrs;
(4) the beginning of stone age: 7 thous. Yrs;
(5) the beginning of copper age: 5 thous. Yrs;
(6) the beginning of iron age: 3 thous. Yrs;
(7) the beginning of atomic age; 1945;
(8) the beginning of cosmic (space) age: 1957;
(9) the beginning of noospheric future: the end of the second millennium.

GEOLOGIC COLUMN FOR THE RIGHT BANK OF THE DNIEPER near the village of Khalepye, Kyiv Region

System	Epoch	Tier	Index	Column	Thickness (m)	Lithology
1	2	3	4	5	6	7
Anthropogene	Holocene		Q_4		1.5	Modern soil
	Neopleistocene		Q_3		12	Continental loess
	Mesopleistocene		Q_2		1-3	Moraine, outwash (loamy sand)
	Eopleistocene		Q_1		4.3	Bouldery loamy sands and loams
Neogene	Pliocene	P	N_2		3.4	Blackclay
					2-5	Red-brown and mottly clays
	Miocene	Poltava	N_1		6	White quartz continental sand
Paleogene	Oligocene	Kharkiv	P		12.1	Quartz and glauconite sand of continental origin
					9	Kyiv sandy

System	Epoch	Tier	Index	Column	Thickness (m)	Lithology
1	2	3	4	5	6	7
						clay. Green-colored
	Eocene	Kyiv	P		22.4	Marly clay of marine origin
		Buchak	P			Yellow-grey quartz sand
		Kaniv	P			Glauconite sand

Questions and assignments

1. What is meant by absolute and relative age of geologic rocks?
2. Which methods are employed to determine absolute and relative age of geologic rocks?
3. What is geochronological (geologic time) scale?
4. How is it constructed?
5. Name the eras in our planet development.
6. Briefly characterize each of the eras.
7. Speak on the development of life in Cenozoic era.
8. Speak on the development of earth surface in Quaternary period.
9. Name the major events of the Quaternary period.

Chapter 6

Genetic Types of Soil Parent Materials

The unconsolidated mass of sedimentary rock in which the soil forms is known under the name of soil's parent material, or soil-forming rock, the later term being in wider use in Ukrainian soil science.

The greater part of the soils in our country were formed in loose quaternary deposits, which, in their turn, were formed as a result of the processes of weathering, transportation of products and their sedimentation on the continental surfaces in the course of 1 to 2 mln years. Quaternary deposits are very different by origin, composition, properties character of lying and, therefore, they are not a uniform medium for the "settlement" of living organisms and soil formation. Parent materials are the material body of a soil. The latter inherits from them its texture, chemical and mineralogical composition, as well as its physical and physic-chemical properties.

Soil forming rocks affect the direction and rate of pedogenic processes as well as formation of soil fertility, composition, properties, and their water, aeration and thermal regimes, as well as content of nutrient elements and use in agricultural production. A close connection between parent material and a soil is particularly strong at the early stages of soil development. But even later on this connection is not lost, though it becomes less distinct owing to the transformation of parent material in the courses of weathering and soil formation.

Parent materials of soils are classified by genetic principle, according to which the following groups of parent materials were identified: eluvial, colluvial, deluvial, proluvial, alluvial, glacial, glaciofluvial, eolian, lacustrine organic and marine.

6.1. Eluvial Parent Materials

Eluvium are the products of weathering left in the place of weathering of primary rocks, not transported or deposited. Very often it is a mixture of angular unsorted fragments which gradually transit into a solid (primary rock), the latter being very often cracked into large blocks. These blocks are covered with smaller and smaller fragments. As it lies in relatively undisturbed position, eluvium is often called mantle rock. All over its thickness, eluvium is nonuniform by texture and skeletal composition. It may be of different chemical and mineralogical composition. Its particle-size distribution may vary from stones and boulders to silts and clays. It may be carbonatic or leached from lime and acid or saline to varying extent. But the connection with underlying solid rock remains quite distinct.

Eluvium lies on the positive elements of topography, mainly in mountain regions. Sometimes it covers the tops of the mountains and the upper parts of the slopes. Occasionally it may be found on the plains the surface of which was not covered by quaternary deposits. Its composition is dependent on the properties of a primary rock in which it formed, climatic conditions, and the character of land forms, which may affect its thickness. With great activity of ground water and durable elluvial process, its thickness may reach tens of meters and in some deep crevices – even hundreds of meters. Actually in such cases we can speak about the formation of a cortex of weathering which is a regularly "constructed" eluvial profile which develops for a long time in the initial rock as a result of hypergeneous transformation of its material.

For many geologists, the understanding of eluvium coincides with the understanding of the cortex of weathering (B. B. Polynov, I. I. Ginzburg, et al.). In arid climates of deserts and on young surfaces, the eluvium is underlain by clastic cortex of weathering. These are coarse, stony fragments, generally the products of physical weathering, so that fine earth fraction is practically never present in them, as it is transferred downward by percolating water or carried away by the wind. Mineral and chemical composition of such an eluvium is very much the same as that of the primary rock. Soil cover is absent on such eluvium, although there may be some primitive soil formation under cover of certain plant associations hardy enough to survive there.

Soil formation goes on very slowly on coarse clastic sediments. The soils formed are thin-solumed, mainly sod soils. The horizon of soil organic matter accumulation gradually passes into parent material. Very often the soils formed in this way are characterized by high infiltration rate and total porosity but their water-holding capacity is low as well as cation exchange and other forms of adsorption capacity.

In semiarid and humid climate, on solid crystalline rocks, the thickness of eluvium is 1-2 meters and more. Cortex of weathering of eluvium is different from that of the primary rock. It contains less Na, Ca, Si, Fe, Mg and K.

In humid subtropical and tropical climates, beginning with the period before the quaternary period, the elluvial deposits formed may reach 10 and more meters in thickness. It very often is of clay texture and soil profile formed in such deposits may have the lower horizons representing the cortex of weathering. Climatic conditions determine the chemical composition of eluvium which is connected with the stages of prolonged weathering and different migration rate of chemical elements.

Eluvium formation is of great importance for the life on Earth as it functions as a source of all the other sedimentary deposits, both continental and marine. As soil-forming rocks, the eluvial deposits of tertiary period are more widespread than those of quaternary period. They are represented by the eluvium of slates, marls, sandstones, chalk, andesites, basaltes, etc.

Eluvium and elavium-deluviam of slates and sandstones are the major soil-forming rocks in the Carpathians. They do not contain carbonates and are light-textured, from sands to medium loams. They contain the skeleton of solid rocks which increases their infiltration rate. In Subcarpathian and Transcarpathian parts of Carpathian region, much more widespread are the alluvial deposits of ancient origin and of mainly loamy texture.

Acid brown forest soils were formed on Carpathian slopes, whereas in parent materials enriched in calcium and magnesium, sod-brown soils were formed.

On the foothills of the Crimean Mountains the soils were formed in the elavium of limestones, chalk and marl. In the mountain forest zone, they were formed in the eluvium of crystalline igneous rocks like andesites and basalts. On the southern slope of the highest (the southernmost) mountain ridge, in arid subtropical climate, some cinnamon soils formed in red-colored cortex of weathering. The southern slopes of Donets Highland are covered mainly by the eluvium of chalk, marl, and slates in which lithogenic carbonatic Chernozems were formed with high porosity, low water retention and unfavorable water regime and soil reaction for the growth and development of plants.

Some shallow-profiled skeletal soils form in the eluvium of sandstones. It is a coarse-textured parent material. It contains much sand and little clay and does not contain calcium carbonate. Such soils possess unfavorable physical properties and are poor in N, P_2O_5 and K_2O. Their organic matter content is within 2.2-2.5 % and exchangeable magnesium may dominate in soil adsorbing complex. Such soils are skeletal (contain rubble) and susceptible to water erosion.

Slate eluvium played a great role in the formation of soils in Donbass. The slates are varying in mineralogical and chemical composition. The soils formed in them are, therefore, different by their productivity-determining properties. Their sola are short (up to 70 cm); they do not contain $CaCO_3$; their SOM content is low and soil structure unfavorable, as well as other physical properties, owing to skeletality. Rubble content may reach 5-25 % and it increases with depth. Water infiltrability is low which instigates soil erosion. There is an excess of magnesium among the exchangeable cations. Soils formed in limestone or chalk eluvium are less leached and have a slightly alkaline reaction. They are saturated with exchange bases and have a grainy structure. Their SOM content and CEC (cation exchange capacity) are moderately high. They effervesce with addition of 10% HCl. But they are skeletal which lowers their productivity. Eluvium of calcareous and solid crystalline rocks is encountered on the right banks of the rivers.

6.2. Colluvial Rocks

Colluvium (from Latin colluvio = accumulation, mixture) are the deposits formed mainly in mountain regions near the foothills and feet of the mountains in the form of land slips and talus slopes. They move along the slope by the force of gravity. Such phenomena are less often observable on the plains, as the slopes there are gentle and short. On the plains colluvial deposits may be found below the river-valley slopes or on the bottoms of the gullies. It is an "azonal" parent material of soils. Gravity causes motion and plays an important part in erosion. Gravity starts materials moving down slopes toward rivers. The pull of gravity causes loose material to move down a slope. Some mass movements are slow while others are fast. In either case, weathered fragments move from higher to lower elevations. Eventually, these materials come to rest at the bottom of the slope. The material that accumulates at the foot of the hill, or cliff, is called a talus pile. Such piles are recognized most easily in semiarid country. But the process can be seen on a small scale in some ditches. The particles of clastic rocks are very unsorted and stony. Rockfalls may even close a road completely. So on rides through mountains, you become familiar with the sign, "Beware of Falling Rock". Any process that makes a slope steeper also brings about mass movement. Sometimes slopes are steepened because they are undercut by waves of rivers. There are many forms of mass movement. Landslides are rapid movements of large amounts of material. Landslides = зсуви in Ukrainian. Sometimes barge blocks break away from steep mountain slopes. These masses then move down the slope as a unit or break up into smaller masses. Landslides often follow long periods of rain.

Mudflows (селі, сельові потоки) also are rapid movements. These flows occur after heavy rains. The mudflow follows old channels down to the valley below. There the thick mud spread out into a cone-shaped mass. Creep (солефлюкція) is common in areas where periods of freezing and thawing alternate. Creep occurs in humid regions on slopes covered by vegetation. Slump (сповзання) is when resistant rock lies on a weak layer. The weak layer, usually shale, tends to slip down the slope. It carries the resistant rock along as a unit. Slump may leave resistant rock along as a unit. Slump may leave a curved scar in the higher elevations where the resistant layer has pulled away (a hillside scar). As a soil forming rock all these kinds of colluvium are practically nonexistent as they constantly get mixed. They make an initial material for deluvium, alluvium, proluvium, marine and lacustrine deposits.

6.3. Deluvial Rocks

Deluvium (from Latin deluvio = to wash off) is the product of weathering of rocks, transported and redeposited along the slopes and especially in their lower parts by nonturbulent rain or melt water. Deluvial deposits are the former eluvial ones transported and deposited along the slope by surface runoff. Such sediments are layered and their particles are well sorted. Their depth increases downslope. Sometimes the layering of their profiles is the inverted layering of the respective upland profiles.

The intensity of washing off depends on climate. The most intensive it is in humid and semihumid climates on plowland. In arid regions it is very weak. Deluvial deposits may be nonuniform by texture. Coarse material may lie on the foot of the slope, especially if the slopes are made of solid rock and the slopes are gentle. Such deluvium is typical for arid zones with the dominance of physical weathering. At the feet of the slopes, the deposits are of different texture: loamysands, loams, and clays as kinetic energy of bedless flows is low.

In lower portions of the slopes the thickness of deluvial deposits may reach 5 or even 20 meters, whereas the length of deposition may be from a few meters to 1-2 kms. In such cases, the deluvium is sorted in such a way that the coarser material lies closer to the top and the finer material is deposited further down the slope, where its clayey texture is no rarity.

One of deluvium’s characteristic features is the absence of distinct layering along the slope and its lying at the feet of gentle slopes, which are the most widespread on plainland. The particles are slightly rounded and may contain plant residues and the remains of shells, as well as the fragments of bricks and soil organic matter. It lies mot on the rock from which it formed but on some other rock. So it may differ very much from the underlying rock material in both texture and mineral composition.

Typical deluvial shelves are absent at the feet of steep slopes, as gravitation processes dominate on such slopes and form colluvial deposits. Or rather colluvial-deluvial ones.

There are two facies of deluvium – that of slopes and that of gullies. Slope deluvium lies on the lower portions of the slopes, whereas gully deluvium forms the cones of discharge in the gully mouths.

Loose deluvial deposits are very widespread and serve parent materials of soils which isolate the soils from the compacted rocks below them.

6.4. Proluvial Rocks

They were described earlier as mudflows. They are the products of weathering transported very rapidly by temporary flows of great intensity along the old valleys. Cones of discharge are the main form of proluvial sediments. Sometimes they merge with each other forming a proluvial plain. Such plains are called concave plains, with specific concentric zonality, being sorted in space. In the upper parts of discharge cones, proluvium is composed mainly by coarse clastic material or pebbles. Farther from upper parts the particles become finer, loamy sand or even clay. The material is only slightly sorted and slightly rounded. Soil-forming properties of proluvium are determined by its texture. The finer the texture, the more productive soil forms in it.

6.5. Alluvial Deposits

Deposits left by flowing waters are called alluvium. Flowing waters in this case is the river waters. Rivers are the most important erosional agents because they affect so much area. Landscapes are largely the work of running water, even in desert regions. Runoff erodes the land about 0.3 meters in 9000 years. One fourth of this sediment reaches the ocean. The rest remains within the drainage basin. A small decrease in velocity causes a stream to deposits its load. Sediment may be dropped along the river sides, within the channel, or at the mouth. The material deposited is well sorted by grain size. The valleys of ancient and present-day rives are covered by alluvial deposits formed from the material of disrupted banks, and of the particles carried by river water from the origin and slopes of the valley. The material is not only sorted but smoothed by running water. Many geologists note that inclined (sloping) layering is one of the diagnostic features of alluvial deposits, as well as rapid and considerable changes in the spread and thickness of deposits, roundness of particles, freshwater fauna remains are lying in the river valley in the form of layers (banks) up to 20-60 meters thick. Alluvial deposits occur in floodplains, terraces, fans, piedmonts, and deltas. Alluvium within the floodplains is divided by its lithological composition into three facial kinds: riverbed, floodplain, and that of ox-bow lakes.

Riverbed alluvium forms in the process of riverbed formation and its evolution towards the profile of equilibrium. The flow ruins the banks as well as the bottom of a riverbed. The valleys very often are much wider than the river channel. The meanders of this channel show in an indirect way the vigor of running water. A lot of loose material is carried away and deposited within the riverbed. In the valleys of plainland rivers, alluvium is composed of sand particles with occasional admixture of gravel or pebbles. Sand particles are of different size well washed off by the flow of water. They do not contain silt or clay and display an inclined layering. Sand grains are well smoothed, of whitish color. When dry, alluvial sands are easily moved by the wind. Their infiltration capacity is high, whereas water-holding capacity and capillary rise are low. They do not contain soluble salts and their cation exchange capacity (CEC) is very low. Low-productivity soils form in such parent material, mainly sod or sod gleyed soils: As parent material of soils, such alluvial deposits cover "borovi" (with pine forests growing along them) terraces of river valleys (ancient alluvium).

Floodplain alluvium is a material deposited by a river during the floods. Such alluvium has a distinct layered profile, the layers being of small thickness and nearly horizontal. It may contain different inclusions of humus material, weed seeds, plant residues and even small brick fragments. It also may contain some newly-formed salts and other minerals (gypsum, carbonates, limonite, marl, etc.).

By mechanical composition floodplain alluvium may be loamy sand, loam and sandy loam and it covers floodplains as an undisturbed mantle. The amount and composition of floodplain alluvium in a given river basin is dependent on climate, vegetation, etc.

Ox-bow-lake alluvium (старичний алювій) are deposits of lacustrine-deposit type because they form in ox bow lakes – the remains of old riverbeds. Such lakes are filled with standing water, so some fine particles carried by muddy water. It is composed of fine-textured sands, loamy sands, silt and clay and are of a distinct thin layered character. Some layers may contain the remains of different swamp vegetation, algae, etc. Sometimes a layer of peat may cover the surface of such lacustrine deposits.

By their age, the alluvial deposits may be recent and ancient. Ancient alluvium lies on "borovi" terraces and recent alluvium lies within the floodplain of a river valley. Alluvium on the terraces as well as alluvium within the floodplains are quite widespread parent materials in Ukraine. Some quite productive soils may form in flood plain alluvium. It is enough to remember the historical productivity of Egyptian soils in the valley of the Nile, subjected to annual floods, which played such an important role for the development of civilization in ancient Egypt. The process of soil formation in floodplain material has such a specific feature that the profile (solum) constantly grows up. The containuous addition of younger layers on the top of older ones is a factor slowing down eluvial processes.

6.6. Glacial Deposits

Glaciers perform a tremendous ruining, transporting and accumulating function on the earth's surface. Moving glaciers carry a great amount of clastic material and deposit it. Such clastic material is called moraine. There are three types of moraines: bottom, ablation and terminal. Bottom moraines are the most widespread and they form vast mantles of cover from a few centimeters to tens of meters in thickness. Ablation moraine forms during glacier melting and deposition of material formerly frozen into ice along the glacier path. Ablation moraine is also known as glacial drift. There are two types of drift – till and outwash, the latter being transported by glacial melt waters and, therefore, slightly sorted. Coarse clastic material dominates in drift.

Terminal moraines, or end moraines are the ridges of clastic (boulders, pebbles, rubble, etc.) accumulated in front of a glacier. Or it may be a mound of tile deposited by retreating melting glaciers.

By texture the moraines are mostly loames with boulders, sandy loams and clays being encountered less frequently. They are always skeletal and contain the boulders of solid rocks carried from afar. Coarse clastic content gradually becomes lower in southern direction. Moraine texture determines its chemical composition. Boulders and pebbles are the fragments of solid rocks, sand and silt fractions are represented by the grains of certain minerals of which quartz is up to 80 %, feldspar – up to 20 %, and micas – up to 4 % by weight. Feldspars dominate in silt and quartz – in sand fraction. Clay fraction may contain amphiboles, granates, hydrous micas as well as admixtures of Kaolinite and montmorillonite.

By their chemical composition, the moraines may be carbonatic (containing lime) and noncarbonatic (leached from lime). The more widespread are acid moraines which formed predominantly from granites and gneisses. Podzolic and sod-podzolic soils form in such moraine. In moraines developed from the slates, rich in organic matter, the soils form which are richer in SOM, sesquioxides and other components which make them look like Chernozems. They are very fertile and totally employed as a farmland.

Carbonatic moraines impede podzolic process of soil formation as they neutralize the acid products of weathering and soil formation, including the decomposition of plant residues. Calcium in parent materials makes the bacteria more active in residue decomposition. Calcium and magnesium carbonates flocculate organic and mineral soil colloids and retain them in the soil. Sod-carbonatic soils (Rendzinas) often form in such moraines. They are relatively rich in SOM of humatic, not fulvatic types. Soil reaction is close to neutral or slightly alkaline, and content of macro – and microelements of plant nutrition may be considerable. The water regime is not so good, as their sola are thin and skeletal (stony).

The moraines are characterized by high content of skeletal particles, they are unsorted or very poorly sorted; their color is a reflection of their petrological composition, and the extent of oxidation of their components, primarily the compounds of iron. Moraine clays and loams possess high values of particle (solid phase) density.

6.7. Glaciofluvial Deposits

Stratified deposits or outwash are rock materials deposited by streams of glacial meltwater. Known as glaciofluvial deposits, this stream-sorted material may produce a variety of land forms. Glacial meltwaters formed a system of streams or flows which sort out the material of a moraine so that the particles become rounded and smooth.

Coarse fragments are the sands with gravel, pebbles, etc. They are deposited near the terminal moraines. Further on, on large areas, plains are formed of well-sorted sands with finer and finer texture. Such deposits create sand fields. Prypyat Polissya may serve an example of such fields. Such sediments are characterized by distinct layering, sortedness and leachedness from the carbonates. They do not contain boulders. The thickness of glaciofluvial deposits vary from 1-2 to 20 meters. Soils formed in such deposits are characterized by low fertility. Their OM content is low. They are poor in nutrient elements. Their water-holding and cation exchange capacities are low.

Banded clays also belong to glaciofluvial deposits. They form in glacial lakes. They have many-layered profiles with layers 0.5 to 1.5 millimeters thick. They are annual accumulations. If such clays are not saline the soils formed in them are quite productive.

Covering loams or overlying loams, as they might be called, are widespread in the areas of glacial deposits. Their origin remains problematic. Some say that they are the product of physical weathering of moraines in cold climate. According to D. K. Glinka, covering loams are glacial outwash. Covering loams are composed of the material properly rubbed by ice and redeposited by the flows of meltwater. They are more widespread in European Russia and as a kind of a mantle they cover the moraine. By texture they are silty loams which contain up to 55 % of coarse silt fraction. Quartz is a dominant mineral of covering loams. Clay content is within 10-35 % by weight. Calcium is a dominant exchange cation in the cation exchange capacity. They are less porous than loess deposits, but more capillary active and with lesser infiltration capacity. Podzolic, sod-podzolic, and grey forest soils formed in covering loams. Such soils are known for their slightly excessive wetness.

6.8. Eolian Deposits

Continental deposits formed by the action of the wind include two very widespread forms – eolian sands and loess.

Eolian sands are mainly the deposits of the rivers, lakes, seas, and eluvium, transferred and redeposited by the wind. The sands are very fine and fine by texture, with particles 0.05-0.25 millimeters in diameter reaching 80-99 % by weight, whereas silt particles are nearly always absent in them. Sand grains are well rounded and smooth of yellow to yellow-red color. They lie in inclined layers. In deserts the thickness of such deposits is the greatest reaching several dozen meters.

Mineralogical composition of sands is varied and depends on the rock from which they formed, as well as the conditions in which the weathering occurred and the way of transportation. Eolian sands are mono- and polymineral. Monomineral sands are 98-100 % a definite mineral, which is quartz in the majority of cases. Quartz sands are very resistant to further weathering. Alluvial deposits in the valleys of the Don, Dnieper, and Volga as well as in those of their tributaries consist mainly of such sands, which undergo transfer and redeposition by the wind.

Polymineral sands are young. They lie at the feet of the young mountains, formed as a result of Alpian orogenesis. They contain no more than 90 % of quartz, feldspars, micas, glauconite and other minerals, which are not as resistant to weathering as quartz. Mineralogical composition of eolian sand determines their chemical composition and, therefore, the fertility of the soils formed in them. Such soils have a very low cation exchange capacity and high red-ox potential, which impedes the processes of soils organic matter formation owing to an excessive oxidation (mineralization) of organic residues. Low humus content in sandy soils is also related to low content of colloidal matter. Such soils have low height of capillary rise and high infiltrative capacity. Very often they do not retain enough water for agricultural crops to be grown on them. But there may be enough water for the pines and other forest trees. The sands, devoid of vegetation, are very movable, and are continuously redeposited by the wind. The wind in this case plays a role of a secondary agent of sediment formation as the prime factor of sand origin may be deluvial, alluvial, proluvial, glaciofluvial, marine, or volcanic processes. In the Forest, Forest-Steppe, and even Steppe zones of Ukraine, some slightly podzolic soils may be formed in such parent material, or sod soils known in Ukrainian as “дернові борові грунта”. The productivity for field crops is very low. It is best to leave them under natural vegetation.

Loess and loess-like loams. They say that the term loess is of German origin and means “not solid, loose”, “weak”, “fluffy”, etc. They claim that term to be proposed by a German scientist K. Leonard in 1823.

Loess is a rock of problematic origin. The enigma of its origin is not yet penetrated though they try settle this question for already nearly two hundred years. One M. I. Kriger detected over 20 hypotheses of loess origin. Be it as it may, but Ukrainian geologist P. A. Tutkowski proposed a well-grounded theory of eolian origin of loess. He thought that the wind transferred silt material from the surface of the glacial deposits. Most researches which study Ukrainian loess admit its eolian origin (A. M. Nabokikh, V. I. Krokos, G. G. Makhov and V. V. Dokuchaev). Present-day international science gives proves of the eolian origin of loess. For good understanding of land forms and soils in steppe regions, it is important to understand the relation between Ice Age aridity and loess deposition. Loess is well-sorted, usually calcareous, non-stratified, yellow-grey eolian clastic sediment. It consists predominantly of silt-sized particles and contains normally less than 15 % sand. It covers the land surface as a blanket which is less than 8 meters thick in the Netherlands but can reach up to 40 meters in Eastern Europe and 330 meters in China. Loess is very porous material and vertical walls remain remarkably stable. But loess slakes easily and is prone to water erosion.

The loess material is produced by abrasion of rock surfaces by glaciers and flown out from glacial outwash plains and alluvium. Deposition is still going on today, at a rate of several millimeters per year. Long-distance transporta of dust particles from the Gobi desert seems to be responsible for the thick Chinese loess deposits. The eolian origin of loess is evidenced by the following facts:

(1) loess occurs as a blanket over a wide range of surfaces, to a large extent independent of topography;

(2) loess blankets are thicker on the leeward (підвітровий) side of obstacles;

(3) there is absolutely no correlation between the mineralogy of the loess blanket and that of subsurface strata. This rules out the possibility of in-situ weathering;

(4) the grain-size distribution of loess is typical of material transported in suspension;

(5) grain sizes show a downwind fining gradient, away from the source;

(6) loess deposits become thinner away from their presumed source;

(7) fossil terrestrial snails have been found in loess deposits;

(8) intercalated (layered) paleosols (soil fossils) are common in most loess belts, and

(9) loess deposition still happens around desert areas today.

In mineral composition of loess, primary minerals are represented by feldspars, micas, quartz, pyroxenes and amphiboles; secondary minerals – by clay minerals, carbonates, sulphates, phosphates and oxides and hydroxides of iron, silicon, and aluminum. Loess may contain up to 0.5 % of organic matter. Coarse silt (0.05-0.01 mm in size) content is loess may exceed 50 % by weight. The calcium carbonate percentage may be up to 30 % of loess weight. Chemical composition of Ukrainian loesses are given in the following table (Table 14).

Table 14. Chemical Composition of Ukrainian Loesses (% of dry weight)

Components	Kyiv Region	Kherson Region
SiO_2	73	66
Al_2O_3	6.9	15.2
Fe_2O_3	2.7	6.5
MgO	1.3	2.8
CaO	6.5	4.0
K_2O	1.6	2.1
Na_2O	1.0	1.1

Phosphorus and nitrogen contents in loess are seldom over 0.5 and 0.01 %, respectively. Material texture is practically the same all over its vertical column. Texture usually becomes finer in southern direction. There are numerous micropores in loess material which makes its total porosity to be up to 50-60 % by volume, and bulk density – within 1-2 g/cm^3. Natural inclination (sloping) angle for loess cliffs is up to 90 %; so they are easily disrupted. Loess is prone to subsidence on wetting drying and that is why we have so many saucer-like depressions (блюдця) on the surface of a loess plateau. Loess, rich in coarse silt, is too susceptible to erosion by gullying. That is why we have so much of a badland within the Slovechno-Ovruch loess upland (elevated plateau). Gully erosion is relatively common in loess materials. The nearly vertical sides and flat bottoms of such gullies are described as U-shaped. Most other gullies are V-shaped. When an exposed loess surface becomes saturated with water, it is subject to mass movement ranging from small slumps to large landslides.

But in other respects loess is an ideal soil parent material with good physical properties and favorable chemical composition.

Loess-like loams are the result of loess transport and deposition by water and, therefore, layered in structure. The lower landforms within loess plateau are covered by them.

6.9. Lacustrine Deposits

They are the result of the geologic work of the lakes. This geologic work is weak in respect of erosivity and quite intensive in respect of product accumulation. The work of lakes is very much dependent on climate and hydrologic regime, as well as on the area and capacity of the inland water reservoir – the lake. The lakelaid clays are the most widespread of lacustrine deposits. They are composed mainly of fine sediment carried past deltas into the main body of water. The texture of the deposit may vary from silt or sand during seasons of high runoff to fine clay during cold or dry seasons, thus producing a layered effect. These layers are called verves and are sometimes counted to estimate the age of the deposit.

Glaciers, crossing hilly topography, produced many small lakes whose locations are marked by lacustrine deposits. They are also called glacial lacustrine deposits by the scientists in Ukraine and are quite common in the Forest zone (Polissya). In arid regions the lakes may be saline. Evaporation of water from the lake causes the formation of oversaturated solutions and soluble salt precipitation. The first to precipitate are the carbonates (calcite and dolomite), then – the sulfates (gypsum, anhydrite, and mirabilite) and finely, if NaCl gets into lake water, a halite sediment is formed. Soda lakes arise as a result of solonchak solodization (solonchak → solonets → solod). The deposits of such lakes contain very little organic matter and practically do not contain sapropel silt. They, therefore, are of light color.

When the bowl of the lake is filled with sediments, an accumulative plain forms. Lacustrine deposits may contain the fossils of fresh water fauna. If lacustrine deposits contain coarse clastic material the soils formed in them are of low fertility. But the soils formed in sapropel-containing sedimentary rocks are mainly swamp soils. The soils in nonsaline lacustrine clays are quite productive. Lacustrine deposits may be valuable as raw material for different industries including the manufacture of fertilizers.

6.10. Marine Deposits

Quaternary marine deposits were formed as a result of geologic work of the seas. Lateral zone deposits are called terrigenic (of Latin terra = land). They are represented by clastic and organic material. On the beaches the eroded material is deposited in the form of well sorted clastic deposits (boulders, pebbles and gravel). Beaches are found on coastlines where eroded material in the sea has been deposited. Beach fragment size depends on local rock type and wave energy. On the continental shelf (part of the continent covered by seawater) of shallow seas the deposits formed are sands with well-expressed layering. They are well sorted bythe size of particles. On the continental slope the deposits are of mixed clays and silts rich in iron oxides, carbonates and organic matter. The deposits in deltas are represented by sandy silts that come from the rivers. They may contain organic matter and the new formations of phosphorites and iron-manganese beans.

Marine deposits are encountered in Ukraine on Black Sea Lowland, on Black Sea coast and its bays, on the spits and in Sivash region. Such deposits are always saline and saline soils form in them.

6.11. Parent Materials of Ukrainian Soils

6.11.1. Parent Materials of the Forest Zone

The most common parent materials of the forest zone are glacial till, glacial outwash, and alluvial deposits. Less common are loess and loess-like loams, peats, lacustrine, and other deposits. Ukrainian Polissya is a zone of mixed forests stretching from the west to the east for 750 kms, and from the north to the South – for 180 kms. Its area is of 11.4 mln hectares (19 % of the total area of Ukraine). Glaciofluvial deposits are the predominant parent material here, occurring on the area which is 10.5 % of the total area of Ukraine. The sand is coarse and medium-grained and composed of quartz with rare admixtures of feldspars or hornblende. Sometimes small pebble and gravel may be found in their layers. Glaciofluvial loams are very rare. The color of glacial outwash is all shades of yellow to yellowish grey. Very often they are underlain by moraine, igneous and hard sedimentary rocks like limestones and chalk. The most common in Polissya sod-podzolic soils (albehuvisols, spodosols) were formed in glacial outwash. The soils are sands cover the plains with predominantly level topography. On elevated elements of topography (Mounds, ridges, and small hills), the parent material of soils is represented by glacial till.

Glacial till deposits (moraine) are found as small “islands” on the elevated spots. These “islands” are surrounded by the “sea” of glacial outwash. More often than not, moraine serves not a parent material but as an underlying sediment covered by glaciofluvial sands and loamy sands. Till soils are better supplied with water and nutrients than their outwash counterparts. More productive soils form in lime-containing moraines. But such moraines are much less common than leached ones. Light-grey and grey forest soils usually form in lime-containing moraines. Those are Forest-Steppe, not Polissya soils, although they form within the borders of Polissya. The same quite productive soils form in loess or loess-like loams of the "island loess plateaux" within the Polissya zone.

Alluvial deposits occupy about 9 % of the total area of Ukraine. They are intrazonal, unlike moraine and glacial outwash.

Recent alluvium was deposited by flood water along the present streams in recent time. Alluvium is layered, predominantly sandy in texture with sand grains well washed off the Fe_2O_3 and other coatings. Ancient alluvium on the terraces of the river valleys is predominantly coarse sand in texture so that the soils formed in it are good for pine forests and no good for field crops because of their low water-holding capacity and poor fertility. If not sufficiently covered by the plants, such sands are very prone to wind erosion. So the ripples and small dunes sometimes embellish their surface.

Eluvium of hard calcareous rocks occupies 2.3 % of the total area of Ukraine. Limestones, chalk, dolomites, marl and other sediments may be found among them. Eluvium of chalky marl containing 50 to 100 % $CaCO_3$ may be encountered in the west and south-west of Polissya. Ukrainian rendzinas were formed in this parent material. They are stony and rich in $CaCO_3$, with neutral to slightly alkaline reaction which makes unsuitable for potatoes, flax, and some other Polissian crops, favoring slightly acid soil reaction. But such soils are moderately good for wheat, rye, and sugar beet. Apple orchards may suffer from iron deficiency.

Kaolin clays may be encountered as parent material in Zhitomir and Rivne Polissya, being the product of granite (feldspar) weathering. Polissya Kaolin formed on the spots of crystalline shield does not contain $CaCO_3$ as may the Kaolins in the Forest-Steppe and Steppe zones.

Organic parent material (peat) was formed in wet areas where plant remains did not decompose but remained around the edges of the lakes. The lakes were eventually filled with organic material and developed into areas of muck and other organic deposits. Central parts of the flood plains are sometimes covered with peat. Lowland peats dominate in Ukraine, which are richer than upland peats. As parent material, peat may be acid, neutral, and alkaline. Some of them are calcareous.

Most of them are not. On the left bank of the Dnieper in the lowland province of the Forest zone, the deposits of peat may be saline.

6.11.2. Parent Materials of the Forest-Steppe Zone

The Forest-Steppe Zone of Ukraine forms a broad belt to the south of the Forest zone, 1500 kms in length and 250-350 km in width. Its area is about 20.1 mln hectares, which amounts to 33.6 % of the total area of Ukraine.

Loess and loess-like sediments of loamy to clayey texture are the dominant parent materials of soils in the zone. With the exception of floodplain sandy soils, light-textured soils on the terraces, severely eroded or xeromorphic soils on the slopes and the soils formed in the eluvium of igneous and hard sedimentary rocks, the bulk of the soils in the Forest-Steppe zone were formed in loess and loess-like redeposited loams.

Loess in Ukraine is a parent material of soils on 74.8 % of its total area. Sand and silt loams are the most abundant, but to the south and east the texture of loess becomes heavier and loamy clays and clay loams appear. As a rule, these parent materials are rich in calcium carbonate. They are suitable for the accumulation of soil organic matter and formation of what is called in our country agronomically valuable structure.

Though in many respects an ideal parent material, loess and soils formed in it may be susceptible to erosion caused by both water and wind. Gully erosion in the northern Forest-Steppe zone as well as on the “loess islands” of Polissya, sometimes makes quite an ominous spectacle to see. The fraction of coarse silt (0.01-0.05 mm in size), which amounts to over 50 % of the weight of local loess is to a great extent responsible for the growth of gullies. Stream bank erosion (the removal of soil material from the sides of running streams) is usual for the right bank of the Dnieper and other rivers of Ukraine. Erosion damage to the ditches along the roadways may be encountered too.

A “single-loess” terrace on the left bank of the Dnieper is covered by a saline variety of loess. Sodium bicarbonate and even carbonate (which drastically increases soils alkalinity) are fount among the soluble salts in local soils.

Typical Chernozems (Mollisol, Chernozem Haplic, Chernozem Chernic) dominate over entire Forest-Steppe zone. They occupy the leveled plateaux with slow runoff and some higher loess terraces. Grey forest soils dominate on elevated well-drained watershed plateaux. They spread on the well-drained watersheds mainly on the right-bank part of the zone with Podilska and Prydniprovska highlands. Loess dominates here as parent material. Consolidated clays, alluvial deposits of sand and some other materials represent the other, much less common sedimentary parent materials.

6.11.3. Parent Materials of the Steppe Zone

The Steppe zone of Ukraine is divided into two zones: Chernozemic Steppe zone and Arid Steppe zone. Chernozemic Steppe zone occupies nearly all of a southern half of the state and forms a wide belt up to 500 kms wide and 1100 kms long. The entire Steppe zone occupies an area of 25 mln hectares, which exceeds 40 % of the total area of Ukraine. Chernozemic Steppe zone exceeds 20 million hectares and Arid Steppe zone reaches 4.7 million hectares.

Loesses are the dominant parent materials in the zone. They are dominantly heavy, clay-like loams. Their texture becomes finer from the north to the south. In the same direction they increase their calcium carbonate content and salinity reduce the depth of gypsum accumulation in soil profile.

The topography of Chernozemic Steppe zone is prevailingly level, although in its northern subzone, there are some elevated areas, 150-300 meters above sea level: Bessarabska, Podilska, and Prydniprovska highlands, as well as the south-western slopes of Middle-Russian Highland and Donets Kryazh. On the Donets Kryazh and accasionally on the other highlands, parent materials are represented by eluvium and deluvium of igneous and hard sedimentary rocks, such as sandstones, limestones, slates, chalk, marl, etc. The soils formed in such materials are shallow, ckeletal and susceptible to erosion. Loesses and loess-like loams of the Prychornomorska lowland are salted with sulfates and chlorides. Red-brown clays underlying them and limestones of a tertiary period, underlying the clays may sometimes serve as parent materials on the slopes. But the dominant parent material on the slopes of the ravines and inactive gullies in the Steppe are represented by deluvium and alluvium of loess.

Ordinary Chernozem (Mollisol, Chernozem calcic) is a dominant soil subtype (or type in accordance with present-day Ukrainian classification) in the Northern subzone of the Chernozemic Steppe zone. Parent material (loess) contributed very much to the formation of this “giant” of soils productivity formed in the virgin steppe under grassland vegetation containing the species of meadow sod-forming grasses and more xerophytic species of feather grass, bluegrass and needle grass.

The diversity of soil cover is very much enriched by the lithological series of soils formed in consolidated clays, alluvial deposits of sand, eluvium of calcareous and noncalcareous solid rocks and parent materials with varying extent of salinity.

In the Crimean Steppe, the Southern Chernozems form mosaic patterns of soils combinations with the Chernozems formed in the eluvium of consolidated lime-containing rocks and clays. Gleyed (with the features of reduction processes) loesses make a parent material for the Solot soils of the pods, flat-bottomed depressions in the Southern and Arid Steppe of Ukraine, filled with water during a certain period of the year.

6.11.4. Parent Materials of Mountain Regions

The Carpathian Mountain region is located in the western most part of Ukraine and it includes the whole territory of Zakarpatska, the greater part of Ivano-Frankivska, the south of Lvivska and the west of Chernivetska oblasts.

On the mountain slopes of the Carpathians, acid brown forest soils were formed in eluvium and eluvio-deluvium of sandstones and shales. These rocks are noncalcareous of sandy to loamy texture and skeletal to a varying degree. They contain rounded and angular fragments of solid rocks. Their permeability to water is rather high. In the lower parts of slopes and in the valleys of mountain streams, pebbles and conglomerates may be encountered.

Ancient alluvial deposits of loamy texture may be encountered in the Transcarpathian regions on both sides of the Carpathian range. Alluvial deposits are rich in pebbles and gravel.

Sheltered by the mountains from the north winds, the southern shore of the Crimea has a mild climate of the Mediterranean type. The cortex of weathering here is red-colored. The products of weathering of andesites and basalts, clayey shales, limestones, and other solid rocks are rich in iron and aluminum oxides and hydroxides. Cinnamon and red-cinnamon xerophyte forest soils were formed in these parent materials.

On the mountain slopes, the parent materials are mainly calcareous. On the foothills of the Crimean Mountains the Soils were formed in the eluvium of limestones, chalk and chalky marl. Such parent materials form skeletal soils. The soils on the mountain slopes were formed in eluvium of sandstones and shales. These are brown forest soils with neutral soil reaction which makes them different from intensively leached brown forest soils of the Carpathians.

Questions and Assignments

1. Which rocks can be named parent materials?
2. What genetic types of parent materials do you know?
3. What parent materials are the most common in the Forest zone of Ukraine?
4. Describe glacial till as parent material.

5. Describe glacial outwash as parent material.
6. Characterize alluvial deposits and their role in soil formation in Ukraine.
7. How can you prove the eolian origin of loess?
8. Characterize loess as an ideal parent material of soils.
9. Describe parent materials on Donets Kryazh.
10. What parent materials are the most common in the Steppe zone of Ukraine and how do their composition and properties change geographically?
11. Describe parent materials of soils in mountain regions of Ukraine.
12. Describe ancient alluvium on the terraces as parent material.

Chapter 7

Agronomic Ores

The term “agronomic ores” refers to the rocks and minerals used in agriculture as fertilizers and soil amendments or as raw materials for the production of fertilizers, soil amendments and other materials used in agriculture. The most important agronomic ores are those which contain macro- and microelements of plant nutrition. Soil amendments are also very important as they are used for the neutralization of excessive acidity and alkalinity of soils.

In our country agronomical ores are classified into 8 groups: nitrogen materials, phosphores materials, potassium materials, agronomical lime, gypsum amendments, organic materials, magnesium materials and the materials containing micronutrients.

7.1. Nitrogen Materials

The source of essentially all industrial nitrogen, including fertilizer nitrogen, results from the fixation of atmospheric N_2, according to the following generalized reaction:

$$N_2 + 3H_2 \rightarrow 2NH_3 \text{ proper temperature pressure, and catalysts}$$

Nitrate minerals like sodium ($NaNO_3$) and potash (KNO_3) saltpetres were formerly used as fertilizers. Today as a source of fertilizers they are only of a historical interest, as nowhere in the world are they mined for the purpose of fertilizer production.

7.2. Phosphorus Materials

There are over 150 phosphorus-containing minerals known in the world. But very few of them are used as agronomic ores. Practically all the compounds in the earth’s crust containing phosphorus are the salts of a phosphoric (H_3PO_4) acid.

Apatite [$Ca_5(PO_4)_3(F, Cl, OH)$] contains 42,3 % P_2O_5. 96 % of the total phosphorus contained in the earth’s crust is represented by this mineral. Large deposits of apatite are located on a Cola Peninsular, in Sweden, Finland, east of Tampa (Florida). The sedimentary deposits of phosphorite were formed 10 to 15 million years ago and are buried under 5 to 10 meters of sand or other sediment.

In 1921 Academician O. Ye. Fersman discovered the largest deposits of apatites in the world (Khibiny). Not long ago, Ukrainian apatite province has been discovered. It includes Novopoltava deposits not far from Chernigiv. The primary phosphorus mineral in phosphate rock is fluoroapatite, $Ca_{10}F_2(PO_4)_2$. Apatite is also the source of phosphorus in most soil parent materials.

The acidulation of rock phosphate with sulfuric acid produces **ordinary superphosphate**, as is shown by the following equation:

$$Ca_{10}F_2(PO_4)_6 + 7H_2SO_4 = 3Ca(H_2PO_4)_2 + 7CaSO_4 + 2HF$$

The ordinary superphosphate produced in the preceding reaction consists of about one half monocalcium phosphate and one half calcium sulfate. Hydrogen fluoride is one of the most hazardous pollutants if released into the atmosphere. The gases from manufacture of the phosphorus fertilizer are passed through a scrubber and the fluorite is recovered. Some of the fluorine is used to fluorinate water. Ordinary superphosphate is about 9 % phosphorus, which equals 20 % P_2O_5 equivalent and has a grade of 0-20-0. The phosphorus in monocalcium phosphate is water soluble. Very little 0-20-0 is used today because other phosphorus fertilizers are less expensive on a equivalent phosphorus basis.

Under proper conditions, the reaction of rock phosphate with excess sulfuric acid will produce phosphoric acid (H_3PO_4). Treatment of phosphate rock with phosphoric acid produces superphosphate with a higher content of phosphorus. The reaction is:

$$Ca_2(PO_4)_6F_2 + 14H_3PO_4 = 10Ca(H_2PO_4)_2 + 2HF$$

The same water-soluble monocalcium phosphate is produced, but without $CaSO_4$. The grade is 0-45-0 and has been called concentrated, or triple superphosphate.

Sedimentary phosphate rocks (phosphorites) are widespread in the European part of the former USSR but they are concentrated in two basins: Volzky (Russian Federation) and Dniprovsko-Donetsky (Ukraine). The fragments of phosphate rock may be encountered on the surface or at depth to 20-40 meters. The deposits were formed 130-140 million years ago when a shallow sea covered the territory of Central and Eastern Europe.

In many regions of Ukraine: Kharkiv, Chernigiv, Donetsk, Sumy, Ivano-Frankivsk, Vinnytsa and Khmelnitsky as well as in the Republic of Crimea there are phosphorite deposits and phosphorite meal (ground phosphorite) is produced which contains 16-17 % P_2O_5. Today some organic farmers prefer phosphorite meal as a source of phosphorus. But its application should be restricted to very acid soils which promote its dissolution.

Vivianite [$Fe_3(PO_4)_2*8H_2O$] may contain up to 33 % P_2O_5. Its deposits are connected with peats where it is easily detected because of its bright blue color. It may be common in the lowland (minerotrophic) peats formed due to an excessive wetness caused by the ground water. It is quite common in the peats of Lviv, Volyn, Ternopil, Zhitomir, Kiev, Chernigiv, Sumy, and Kharkiv regions. On the Kerch peninsula it is encountered in the iron ore deposits. Only the peats with over 10 % of P_2O_5 are used as fertilizers.

7.3. Potassium Materials

Potassium mines were opened in Germany in the late nineteenth century. Since then they were opened in many other parts of the world, including Ukraine. These potassium deposits, which were formed when ancient seas evaporated, are called evaporite deposits. The salts in the ocean water precipitated as the water evaporated and the salts in the water became more concentrated. Later, these salt deposits were buried under various kinds of overburdens and rocks.

The minerals in the deposits are representative of the salts in seawater, mainly salts of sodium and potassium. Sylvite is KCl and Sylvinite is a mixture of KCl and NaCl. Langbeinite is a mixture of potassium and magnesium sulfates. Processing of most of the ores consists of separating KCl from the other compounds in the ore; more than 95 percent of the potassium in fertilizers in KCl. Some of the Canadian deposits are relatively pure KCl and require little, if any, processing. After the processing, the material, KCl, is marketed as 0-0-60.

Some KCl is obtained from brine lakes in California, and at the Great Salt Lake in Utah. Potassium chloride is a good fertilizer for the crops which can tolerate chloride anion (Cl^-). These are the grain crops, sugar beet, hemp, cotton and some vegetables. But for potatoes, clover and fruits the fertilizers containing Cl^- should not be recommended.

A very large deposit of sylvite is located near Solikamsk (Russin Federation), on the left bank of the Kama. Prykarpatsky potassium region in Ukraine extends over 300 kilometers from the north-west to the south-east and includes Stebnitske and Kalush-Volynske deposits (basins). Fertilizer named “Kalushit” has a formula $K_2SO_4*CaSO_4*7H_2O$ and contains about 29 % K_2O.

Glauconite is the term used for small greenish earthy pellets (also called glauconite pellets) of heterogeneous mineralogy and for a micaceous mineral. Glauconite contains up to 9.5 % K_2O. It is common in the marine sediments. In Kursk and Orel regions of the Russian Federation glauconite sands underlie the deposits of phosphorites. Glauconite sands are quite common in Ukraine but their use for the manufacture of fertilizer is not economically feasible.

7.4. Liming Materials

They are used for the liming of acid soils. These are limestones, marl, dolomites and some other materials. Agricultural lime is a soil amendment containing calcium carbonate, magnesium carbonate, and other materials, which are used to neutralize excessive soil acidity and to furnish calcium and magnesium for plant growth. Most frequently, agricultural lime is ground limestone that is mainly $CaCO_3$ and is calcitic lime. If the lime contains a significant amount of $MgCO_3$, the lime is dolomitic. Lime hydrolyzes in the soil to form OH^-: $CaCO_3 + H_2O = Ca^{2+} + HCO^-_3 + OH^-$. The hydroxyls neutralize the H^+ in the soil solution and the pH increases.

Organic limestones: coquina, composed of shells and coarse shell fragments; reef limestones, composed of corals and other lime-secreting organisms; chalk, composed of the shells of foraminifers; clayey limestones, containing 5 to 10 % clay are encountered as large accumulations. Chemical limestones, including limestone tufa and pisolithic formations are less common.

Marls are calcareous rocks containing 30 to 50 % of clay and 20 to 75 of carbonetes, and may also be used as liming materials.

7.5. Gypsum Materials

Sodic (Solonetzic) soils have an adsorbed or exchangeable sodium percentage [ESP] of 10 or more. Exchangeable sodium causes the defloculation or dispersion of soil colloids together with a breakdown of the soil structure and appearance of excessive soil alkalinity. The basis for treatment of sodic soils is the replacement of exchangeable sodium with calcium and the conversion of any sodium carbonate into sodium sulfate. Finely ground gypsum ($CaSO_4*2H_2O$) is broadcast and mixed with the soil. The calcium replaces or exchanges for the sodium adsorbed on the negatively charged colloid surfaces. Gypsum, calcium sulfate ($CaSO_4*2H_2O$) is the most common soil amendment used in a ground form. A product of evaporation, it occurs in thickbedded deposits very often together with potassium ores (Solikamsk), near the lakes of Eton and Baskunshak and in other locations.

7.6. Organic Materials

Peats and brown coal (lignite) are the most common of such materials. They are used to increase the organic matter content in some soils. Peat is used to make composts with animal manures which are an excellent source of nutrients for plants. Brown coal is used in the manufacture of sodium humate and other humate preperations which are applied to the soil to increase its humus content and improve its structure and other physical properties.

Questions and Assignments

1. Define the term “agronomic ores”.
2. What nitrogen materials do you know and what is their role in the manufacture of fertilizers?
3. How is apatite transformed into a superphosphate?
4. What is the different between the ordinary and the triple superphosphate?
5. Describe ground phosphate rock as a fertilizer.
6. What potassium materials do you know?
7. Describe glauconite as a raw material for fertilizer manufacture.
8. What is agricultural lime?
9. How does agricultural lime react with the soil?
10. What does gypsum do to a sodic soil?
11. How do organic materials amend the soil?
12. Name the most common organic materials. Describe their use.

Chapter 8

Elements of Geomorphology

Relief is a set of forms of the earth's surface (mountains, plains, hollows, etc.) of different size, structure and origin which are at different stages of development and in complex combination with each other and the surrounding environment.

Geomorphology is the science of land forms, their origin, development and connection with the earth's geologic structure. One of the major tasks of geomorphology is identification and mapping of land forms. Land forms are the features that form landscapes. Landforms include features such as beaches, bars, and spits found along coast lines. They include high mountains and steep canyons found in mountainous areas. The low hills and shallow valleys of the plains and the deep canyons of the high plateaus are all examples of land forms. Features of the earth's crust seem to us to last forever. But they only appear unchanging. However, landforms are constantly changing.

Exogenic forces, exerted by the heat energy of the sun, affect the formation and development of land forms. Geologic work of different factors has already been discussed. The active role of topography among the other elements of geographic environment is connected with its dependence on internal or endogenic forces of the Earth which function in combination with exogenic or external forces (external dynamics). The processes of relief formation and ruination occur in nature at one and the same time. Geomorphology studies landforms in their never ending gradual and rapid changes, for example the river valley formation. Different parts of, for example, Ukraine have landscapes that contain distinctive landforms. There are such large forms as mountains, plains, and plateau areas. Each type of landscape is related to the climate, rock type, structure, and history of the area. So the central place in geomorphology is occupied by the genesis or origin of landforms and their development.

This branch of geological science is based on genetic classification of land forms. They form systems in accordance with the factors (conditions) of their formation and depending on the nature of the motive forces of their transformation. Genetic classification identifies the forms of topography in accordance with the motive forces of their development, taking count of other forces concomitant to them.

In addition to genetic landforms classification, there also exists the morphological classification of relief. It appeared from the interaction of two divisions of geomorphology: morphography, which studies the external features of landforms and morphometry, which studies their guantitative characteristics, like height, size, area, density of placement, slope steepness, etc. Morphographic classification is widely used in mapping and topography.

8.1. General Notions of Land Surface Relief

Relief is a set of flat, concave, and convex forms of land surface. It’s divided into three types: level land, hills, and mountains. Relief type is determined by definite combinations of land forms repeated with certain regularity on the surface of lithosphere, which have similar origin, geologic structure and history. One and the same forms of topography may be encountered in different types of relief.

Plain lands are the sizable areas of dry land with level or slightly wavy surface, as a role, slightly inclined in one direction. Ukrainian plains are vast flat areas that make up about 95 percent of its area (lowlands: about 75 percent and highland: about 25 percent). Lowlands are the plains with absolute elevation not more than 200 meters above nearby lands, whereas the lowlands with absolute elevation over 200 meters above nearby lands, are called uplands.

Hilly topography is an alternation of hills not over 200 meters above sea level with hollows and valleys. These combinations may be of different characters uniformly and nonuniformly combined.

Mountainous topography consists of rock bodies that rise at least 600 meters above nearly lands. This topography includes alternating mountains and valleys. They may occupy vast areas and form mountainous countries.

The majority of landforms by their shape remind us of solid geometry bodies – cones, pyramids, prisms, etc. They have plains, edges, and lines called the elements of topography.

8.1.1. Elements of Topography

Surfaces which form every landform may be differently inclined to the horizon. More often than not their cross-sections are complex. It can be a top, or a bottom, or a slope. The surfaces may be horizontal or inclined. There are no absolutely horizontal and flat surfaces in nature, whereas inclined surfaces are more widespread. On areas close to being level, the swamps, solonchaks and takyrs are formed.

Inclined surfaces may have different steepness. They can be uniformly and nonuniformly inclined. Four features of slope affect velocity and amount of runoff and hence erosiveness: slope gradient, slope length, slope shape, and slope aspect.

Slope gradient (steepness) is measured in units of vertical fall either per single horizontal unit (decimal fraction) or per hundred horizontal units (percent).

Increasing slope gradient increases the speed of water moving downhill and the erosive force of the flowing water. And less water is infiltrated on the slopes compared with level surfaces, which is very important for soil formation: the soils receive less water because of surface runoff. And erosion always increases with slope steepness. For all these reasons, the soils forming on the slopes are less deep.

Slope length is the distance from the crust of a knoll or hill to the point where deposition of transported material starts, or where runoff enters a natural or prepared waterway. Runoff losses per unit area are usually (though not always) the greatest on short slopes. There is generally more erosion on longer slopes in spite on less runoff per unit area. Longer slopes increase the amount of erosion when rainfall intensity is high or the permeability of the soil is low or both.

Slope shape across the slope (along the contour lines) can be straight, concave, or convex, and the downhill direction can also be straight, convex (increasingly steeper gradient downhill) or concave (gradients progressively less steep downhill). Many slopes are convex at the top and concave at the bottom (complex). Convex slopes (in either the contour or the slope direction) cause water to flow away from an area so that the site is drier than it would be otherwise, whereas concave slopes cause water to accumulate and make the soils wetter. Slope-shape effect on runoff and erosion is complex. Runoff velocity is slow and soil movement is minimal near the top of convex slopes because runoff volumes are small and slope is gentle. Water movement is faster lower on the slope because the gradient is steeper and more runoff water accumulates. Soils tend to be shallow on convex slopes; soil replacement from the flatter slope above is slower than loss to steeper slopes below.

Runoff is usually slower on concave slopes because at the top (where slopes are steeper) there is little accumulated water and lower down (where there is lots of water) slopes are flatter. Nearly all natural slopes in humid regions are complex, having convex slopes at the top and concave. The greatest soil losses are observed on the slopes concave across and convex downhill. Complex slopes are observed to lose less soil as a result of erosion than the uniform ones.

Slopes aspect is the direction the slope faces. South- or west-facing slopes in the northern hemisphere are warmer and have higher evaporation during the growing season. Water storage is reduced and there is less plant growth, especially in dry climates. Reduced vegetation usually results in increased erosion. North- and east-facing slopes are usually noticeably cooler, more moist, and better vegetated.

Another element of topography includes the so called characteristic lines which include watersheds, lines of water flow, footlines, and border lines. Watershed is high ground separating two neighboring drainage basins. On one side of it, the water drains in one direction and on the other side it drains the opposite way. The lines of preferential flow are the lines along which the temporary streams run. In Ukrainian they are called капіляри стоку. Footlines outline the bases of slopes belonging to different landforms. There also exist some characteristic points of topography. There are mountain top points, mountain saddle points, river mouth points, and bottom points. Mountain top points have the greatest elevation in a given locality. Mountain saddle points are also called mountain pass points as they are located on the lowest spots of mountain ridges. Mouth points are at the confluence of two rivers or where the river flows into the sea. Bottom points, in contrast to the top points, are located on the bottoms of depressions of topography.

8.1.2. Major Landforms and Their Systems

Mountains. Rock bodies that rise at least 600 meters above nearby lands are called mountains. Dissected mountains are a type of mount. They are not the result of crustal plate collisions. They are formed by the erosion of plains or plateaus. They have a simple structure of horizontal or gently folded layers. They have been cut from the original landscape by erosion. Structural mountains result from uplift within the Earth’s crust. All the great mountain systems of the Earth are of this class. Such systems are hundreds of kilometers wide and thousands of kilometers long. Many of them lie near and roughly parallel to continental coastlines. These mountains are formed by collisions of crustal plates. According to Ukrainian classification, they belong to technical group of landforms. The Appalachians may serve an example. The Rockies too and probably the Urals. They were formed 200-100 million years ago. Structural mountains have several features in common. All have a central area called the mountain core. This core is mostly granitic rock surrounded by metamorphic rock. Metamorphism decreases away from the mountain core. These structures eventually become more gentle and disappear in the flatlands away from the mountains. Fault block mountains also are structural mountains. Great blocks of the Earth’s crust may be raised or lowered during late stages of mountain building. These mountains usually are a part of the great mountain systems of the world. During the uplift of structural mountains, magma flows upward into the crust.

Volcanic Mountains belong to volcanic group of landforms. Volcanoes form volcanic mountains within areas of mountain building and on the ocean floor. Some volcanoes are found on plateau areas next to structural mountains. Lava and other extrusive igneous material from volcanoes build mountains in many areas of the world. The shapes of these mountains are controlled by composition of extruded material. Dome volcanoes form from rhyolite which is more viscous or thicker than andesite. It quickly stops flowing and hardens. Shield volcanoes form from basalt. Basalt lava hardens gradually. Igneous rock bodies are formed within the mountain chain of structural mountains. They were described earlier and are represented by batholiths, stocks, laccoliths, dikes, etc.

Plains are vast flat areas that make up about 95 percent of Ukrainian territory. They make up about one half of the total landscape of the United States. Two examples of their plains are the Atlantic Coastal Plain and the Great Plains. The Atlantic Coastal Plain is the exposed margin of the continental shelf that has emerged from the sea during the past several million years. Wide river valleys have cut through the plain producing low hills between river systems. Much of the area is covered by swamps and lakes. Such forms of topography belong to marine group of landforms, according to our classification.

An example of glacial and fluvial group of landforms finds itself in our Polissian Lowland. The Great Plains also is a large flat region. It is a landscape underlain by sediments that have been eroded from the uplifted Rocky Mountains. Most of the landforms present in this area are related to the drainage patterns developed by river systems. Our Ukrainian Volyn and Podol Uplands are much the same in respect of erosion development including gullying (яроутворення). Like Great Plains, our Prychornomorska Lowland is subjected to deflation (wind erosion). All our plains are rich in erosion and eolian land forms, including the coarse sands of quite humid Polissian Lowland. Many complex drainage patterns have been etched into the loose, soft sediments of the plain.

Plateaus [pla’tɔs] are high areas of flatlands adjacent to mountains. Loess plateaus of eolian origin are very widespread in Ukraine, especially beyond the zone of Polissya. The Appalachian Plateaus in the USA is composed mostly of old, flat-lying sandstone beds that have been deeply cut by erosion. Within the loess plateau of Ukraine, there are numerous shallow depressions (saucers), which form as a result of subsiding owing to the dissolution of $CaCO_3$ and other soluble salts in loess.

Some landforms have biogemic origin (anthills, coral reefs, peat, etc.), where others are the result of the force of gravity (landslides, talus slopes, etc.). Landforms are also divided into positive and negative ones. Hills, knolls, mountains, and mountain ranges are the examples of positive landforms, whereas gully, hollow, valley, and depression are the example of negative landforms. Many topographic landforms indicate what erosional agent is most active in forming the landscape. V-shaped valleys, for example, tell you that rivers are active. V-shaped valleys also indicate that rivers are still cutting their valleys. Over time, as the river erodes its valley, the channel widens. Then the stream begins to meander. When the meander belt is wide, the river is part of a well-developed drainage system. The topography of arid regions is largely the work of runoff and wind. The landforms, however, are very different from those of humid regions. In arid regions, streams flow only after a rainstorm. The river channels are steep-walled.

In Ukraine the sloped are classified by their length: long (over 500 meters), average (50-500 meters), and short (less than 50 meters), and by their steepness in degrees: nearly level (1-3°), gentle (3-6°), slightly steep (6-9°), moderately steep (9-12°), steep (12-15-°), very steep (15-30°), rapidly steep (30-45°), and cliffs (over 45°).

Landforms in our country are divided into types and classes. By types we mean a definite combination of landforms repeatedly occurring on lithosphere's surface can having similar origin, structure, and history. In the practice of field survey it is customary in our country to identify macro-, meso-, micro-, and nanorelief.

Macrorelief pertains to large landforms differing from each other by hundreds and thousands of meters, as, for example, mountain ranges, river valleys, and watersheds.

Mesorelief applies to landforms within macrorelief formations like the hills and hollows within plateaus and lowlands or large river valleys. The difference among the landforms are from a few meters to a few dozen meters. Sand dunes belong to these forms, as well as average hills and knolls.

Microrelief pertains to relatively small landforms, like small hollows or saucer-like depressions in loess plateaux, as well as small hills and knolls of eolian origin within the areas of coarse-textured sands devoid of natural vegetation. The differences in elevations are within a meter or so.

Nanorelief (νάννοσ in Greek meaning dwarf) pertains to certain anthill and molehills, but more often, to the surface of agricultural fields subject to different type of tillage operations. If after plowing we have one kind of nanorelief, then after harrowing we have another one.

Land relief types, by their genesis, are united into classes, as it is shown in the following table (Table 15).

Table 15. Geomorphological Classes and Types (D. M. Katz, 1981)

Class	Type
Accumulative Plains	Forms by ancient glaciations: belts of end terminal) moraines, moraine hills, kames, drumlins, bottom moraine plains, outwash plains, lacastrine-glacial plains, old valleys. Not related to glaciations directly: marine plain, lacustrine-alluvial plains, alluvial plains (plain land and terraces neighboring them, deltas), eolian plains, proluvial deposits (cones of discharge and talus slopes).
Denudation Plains	Abrasive plains, erosive plains, erosive plains on the basis of faults, desert pavements, etc,
Denudated Tectonic Mountains	Mountains after glacial work, Mountains denuded by colluvial processes, avalanches and land slides, flat mountains gullied by erosion, foothills degraded by erosion.
Volcanic Mountains and Plateaux	Landforms of active or dormant volcanoes, which acted in quaternary period, domes, etc.
Anthropogenic Formations	All man-made landforms.

Accumulative plains were formed as a result of geologic work of rivers and temporary flows as well as geologic work of the sea, glaciers, lakes, and wind.

Denudation plains arose on the areas with distinct topography formed by denudation, which includes weathering, erosion, work of glaciers, deflation, etc. and product transportation and deposition. They are now of plain land character. Denudation plains are less widespread compared to accumulative (accumulation) plains.

Denudation-tectonic mountains were formed as a result of interaction of tectonic processes, mainly uplifts, with denudation processes defined above. Beside uplifts, other tectonic processes also led to the origin of such mountains, including landslides, talus slopes formation and like processes.

Volcanic mountains and plateaux, as the name implies, were formed as a result of volcanism.

Anthropogenic formations include all types of topography the origin of which is directly or indirectly connected with the activity of man.

Types of topography unite together different forms of land. Of accumulation plains, the most widespread are river and marine terraces, river deltas, and the cones of discharge, as well as proluvial and deluvial deposits and outwash and lacustrine-glacial plains. The types of relief are closely connected with geologic history of a region and its geologic structure. Specific anthropogenic landforms are encountered in the region of old irrigation, consisting of old irrigation canals and sedimented material obtained in the process of their cleaning. Open drainage systems and artificial land leveling are also the examples of anthropogenic landforms.

8.2. Geomorphological Maps and Their Use in Soil Science, Agrochemistry, and Agronomy

Geomorphological maps are the maps which depict landforms, their genesis, and their age. Such maps are classified by their content, aim, and scale. By the content of geomorphological maps they are divided into special and general:

Special maps reflect special landform characteristics, for example, their morphology, or genesis, or age (morphogenesis);

General maps reflect several landform characteristics, for example, their morphology, and genesis, and age. By their aim or the function they have to fulfill, the maps of landforms must be suitable for the use in the industry, agriculture, tourism, forestry, etc.

The scales of geomorphological maps may be:

Large: > 1:200,000,
Medium: 1:200,000 – 1:100,000, and
Small: <1:100,000.

Morphogenetic and morphographic special large-scale maps are especially useful in agriculture, including soil science, agrochemistry, land surveys and planning, and land reclamation projects.

Soil survey and mapping needs such maps for the issuing of their own maps. Land reclamation needs such maps for the construction of irrigation and drainage systems. The locations of landslides are important for the development of a system of their control practices. Slopes and watershed characteristics are used for modeling and prediction of water erosion events and projecting the practices of soil erosion control and use of land areas in accordance with their capabilities.

8.3. Geomorphological Zoning of Ukraine

In accordance with the geomorphological zoning (P. M. Tsys’, 1962), three subprovinces were identified within the territory of Ukraine:

I. Plain of multiple genesis,

II. Carpathians, and

III.Crimea.

Within the subprovinces, the following regions (oblasts) and subregions (pidoblasts) were identified:

- I. Polygenetic Plain:
  - a. Polissya-Dnipro lowland
    - i. Polissian plainland
    - ii. Left-bank plainland
  - b. Azov Sea-Dnipro upland
    - i. Near-Dnipro upland
    - ii. Zaporizzian plainland
    - iii. Near-Azov-Sea upland
  - c. Volyno-Podil upland
    - i. Volyn upland
    - ii. Internal plains of Upper Bug and Styr
    - iii. Podil upland
  - d. South-Western edge of Middle-Russian Upland
  - e. Azov Sea-Black Sea lowland
  - f. Donets Highland
- II. Carpathians
  - a. Subcarpathians
  - b. External Carpathians
  - c. Verkhovyna Watershed region
  - d. Polonyno-Chornogora region
  - e. Northern border of Marmoros region
  - f. Volcanic Carpathians and the hollows between the ranges
  - g. Transcarpathian plainland
- III. Crimea
  - a. Major Crimea
  - b. Coastal Foothills
  - c. Southern Beach
  - d. Kerch Peninsula
  - e. Bessarab upland

Within the regions and subregions, the districts were identified:

<u>Polissya</u>

1. Central Polissya
2. Kyiv Polissya

3. Volyn Polissy
4. Chernigiv Polissya
5. Novgorod-Siversk Polissya
6. Ovruch Highland
7. Chernobyl glacial deformations
8. Dnipro moraine hills
9. Zamglay Valley

Left-Bank Lowland

1. Upper Terrace of the Dnieper
2. Borova Terrace of the Dnieper
3. Floodplain Terrace of the Dnieper
4. Poltava polygenetic plain
5. Gulyay-Pole Polygenetic Plain
6. Northern-Donets Steppe
7. Orillya Plateau

Right-Bank

1. Kyiv Plateau
2. Kaniv Mountains
3. Near-the-Dnieper Upland
4. Near-the-Bug Plain

Volyn-Podil Plateau

1. Podil Plateau
2. Pokuttya
3. Tovtry
4. Kremenetz Table Hills
5. Roztochchya and Obpillya
6. Dniester-San Plainland
7. Foothills and Subcarpathian Highland

Black Sea Lowland

1. Black Sea Pontic Lowland
2. Nogay Steppe
3. Bessarabian Neogenic Plateau
4. Ekin Terrace
5. Dniester Swampy Floodplain
6. Viurm Delta of the Dnieper
7. Spits and Arches

8. Danube Delta

Donets Highland

1. Mane-like Donbass
2. Donbass of Valleys and Hollows
3. Donets Karst
4. Above-the-Azov-Sea Crystalline Ridge
5. Paleovolcanic above-the-Azov-Sea Plain

Questions and assignments

1. Define the term relief.
2. Which science studies the landforms?
3. Describe genetic classification of landforms.
4. Describe morphographic classification of landforms.
5. What is the type of topography?
6. Name the major types of topography.
7. Name the essential landforms and their systems.
8. Which are positive and which are negative landforms?
9. Characterize landform type systematic (classification).
10. Speak on macro-, meso-, and microforms of topography and their importance for soil formation.
11. What is nanorelief? Give an example of nanorelief.
12. Speak on the use of geomopphological maps.
13. Speak on the geomorphological zoning of Ukraine.
14. Speak on the anthropogenic landforms.

Chapter 9

Human Being As a Geologic Force

Some geologists (for example, I. B. Borgolov, 1989) state, that by its scale and power, the geologic activity of man is now quite commensurate with any of natural geologic forces. By its many-vector action this activity is becoming equal to the combined activity of natural factors, considered earlier in this book. We now need no additional evidence to foresee that human activity on Earth may be even much more destructive. It is enough to remember the stock piling of mass destruction weaponry and a single local Chernobyl disaster.

Before we go any further, we must analyse the dynamics of world population growth. By 1830, the population has reached 1 billion. It took 100 year to add the second billion, 30 years to add the third, and only 15 years to the fourth. Another twelve years and we reached 5 billion. Population growth was accelerating year after year. We heard a lot about demographic explosion in the middle of the 20th century. But explosion is something unpredictable whereas in the course of two centuries (19th and 20th), the human population of the world was growing in accordance with the equation, derived by one of the authors of this text:

$$P_T = \exp\{6*10^{-5}(T-1827)\ |T-1827|\},* \qquad (5)$$

which follows from the differential equation:

$$Dp/dt=Kpt, \qquad (6)$$

where $t=T-1827$, P_T is the population of the world (billions) in the year T (1827, 1800, 1950, 2000, etc.).

Speaking about demographic explosion it would not come amiss to state that during the entire period of accelerated industrial development in the world history (18th, 19th and 20th centuries), the world population growth was an exponential function of the square of time and may be tolerably well described by the following equation (L. R. Petrenko):

$$P_T = 0.867\cdot\exp[4.9*10^{-5}(T-1800)\ |T-1800|]. \qquad (7)$$

The equation is less exact than the for the two centuries (19th and 20th).

According to this equation, the world population in 1700 was:

$$P_{1700} = 0.867\cdot\exp[4.9*10^{-5}(1700-1800)\ |1700-1800|] = 0.53 \text{ billion people.}$$

---

* More exact equation is like that:

$$P_T = \exp\{6.06*10^{-5}(T-1827)\ |T-1827|\}. \qquad (8)$$

In 1800 it was:

$$P_{1800} = 0.867\cdot\exp[4.9*10^{-5}(1800-1800)^2] = 0.87 \text{ billion people,}$$

In 1900 it was:

$$P_{1900} = 0.867\cdot\exp[4.9*10^{-5}(1900-1800)^2] = 1.42 \text{ billion people,}$$

In 2000 it was:

$$P_{2000} = 0.867\cdot\exp[4.9*10^{-5}(2000-1800)^2] = 6.15 \text{ billion people.}$$

A concrete equation may be is not sufficiently exact, but the exponential function of the square of time is, probably, quite valid.

Fortunately for us, humans, the world population growth is now a little bit slower that it used to be in the course of two or three centuries and, according to one of the authors of this text (L. R. Petrenko), in the spell of time between 2000 and 2100, we, at worst, can expect this growth to proceed in accordance with the following equation:

$$P_T = 16\cdot\exp\{-5|2240-T/418|^3\}. \qquad (9)$$

It is now April of 2014 when these lines are being corrected and expected population of the world should be:

$$P_{2014} = 16\cdot\exp\{-5|2240-2014/418|^3\} = 7.2 \text{ billion people.}$$

In the year of 2050, for example, it will be:

$$P_{2050} = 16\cdot\exp\{-5|2240-2050/418|^3\} = 9.8 \text{ billion people.}$$

We shall not make any more predictions which are an ungrateful thing to do but we were proud to discover that in the course of two “industrial” centuries (19th and 20th), the world population growth could be described by a simple equation:

$$Dp/dt=Kpt, \qquad (10)$$

which is slightly different from the universally accepted equation of biological (natural) growth:

$$Dp/dt=Kp, \qquad (11)$$

which, in its turn, proves that the growth was slightly “supernatural” and even the two great wars failed to severely indent the curve of population growth.

Surprisingly, such rapid growth has not occurred simply because people are having more babies. The discovery of more effective means of disease prevention, control, and cures has extended life expectancy. Improved sanitation, stronger public heath laws, and more effective food production techniques have greatly reduced the infant mortality rate. Many more children are surviving to reproductive age.

The rising human population has resulted in more that just problems with overcrowding in the world. Our activities produce by-products that are disrupting the natural environment. Even though population growth is highest in developing nations, the so-called industrialized countries use most of the world's natural resources.

Scientists see three major areas of concern on the Earth: the atmosphere, the oceans, and the tropical forests. Air and water pollution are ongoing problems, and only seem to be getting worse. We constantly hear about acid rain, the greenhouse effect, and depletion of the ozone layer. It seems like there is a new oil spill about once a month, not to mention toxic waste, and the vast amounts of garbage dumped into our oceans each year. The loss of our tropical forests may lead to much more far-reaching problems than extinction of species, such as greenhouse effect (because plants absorb carbon dioxide), and desertification.

9.1. Role of Man in the Transformation of the Earth's Crust

The area of "anthropogenic badland" including the areas destructed by erosion, overgrazing, mining, etc. is about 4.5 million of square kilometers or over 3 % of total terrestrial area. Anthropogenic geological process nowadays proceeds in three directions: 1) mining and extraction from the Earth of useful materials; 2) agricultural activity of man, and 3) engineering practices. Every year about 1.7 billion tons of coal and peat, as well as over1.5 billion tons of oil and over 350 million tons of iron ore are extracted from the Earth's bowels. Chemical composition of the Earth's crust changes, as it loses such elements as copper, lead, zinc, tin, silver, gold, etc. The quickest to diminish are the stocks of carbon accumulated in the Earth's crust as coal, oil, and gas. It is used for fuel, which increases carbon dioxide concentration in the atmosphere. In about 100 years the amount of CO_2 in the air will be twice of what it is now. Air temperature will grow considerably and we shall have a hot house of a planet.

Ground water regime is being disturbed very much. A large city like Moscow is using over 120 million cubic meters of water daily and a much larger city of Mexico – 2.5 times more. Earth's crust subsides in many places because of pumping out of ground water: some areas in Tokyo subside up to 20 centimeters, and some areas in Mexico – up to 30 centimeters every year. Ground water tables lower as a result of drainage of swampy areas as well as a result of mining activity.

Soil tillage is fraught with working out of over 3,000 cubic kilometers of the most fertile layer of the soils. This layer is being enriched with fertilizers, disturbed by plows and other implements. Loose and bare, it is subjected to all kinds of denudation, including accelerated erosion by wind and water. In early thirties, the Americans had their dust bowels over the area of Great Plains, and in mid-fifties the Soviets had their virgin lands plowed and subjected to no less severe dust storms. Irrigation on large areas, if done incorrectly, increases the area of salted and sodic soils. Areas drastically disturbed by mining are an environmental menace because they are a source of large quantities of sediment. Some are strongly acid with toxic heavy metals. Iron and coal tailings often contain iron pyrite, which can cause soil conditions too acid for plant growth. Acid mine waters may contain toxic quantities for copper, iron, aluminum, and manganese. In addition, the mine spoils may contain toxic heavy metals such as cadmium, copper, lead, and zinc. Most mine spoils have steep unstable slopes; some are excessively stony; some are deficient in silt and clay and have very low water-holding capacities. Most are low in fertility. In arid regions, mine spoils are too dry and may be too saline or sodic for satisfactory vegetation establishment. Some mine spoils have been reclaimed and revegetated. A large proportion remains as the mining operation left them-bare, scarred by gullies, useless and unsightly.

Anthropogenic geologic activity is connected with destruction, movement and reformation of rocks. Thus, in India the depth of gold mines reaches 3800 meters, in South Africa – 3950 meters. Some bore holes of large diameter may reach the depth of nearly 15 kilometers.

People depend on having forests, lakes, and seashores available to use for relaxation and recreation.

9.2. Natural Resources

Wherever people live and work, they change the air, water, soil, mineral resources, and sometimes even the landforms. But we cannot separate people from the environment. Materials form the Earth’s crust that we can use are called natural resources. Some of the Earth’s resources are renewable, others are not. Renewable resources include forests, soils, water, air, and land. We should use them without losing them for future generations. Nonrenewable resources include metal and nonmetal ores, coal, oil, gas, etc. As the population of the Earth grows, there is greater demand for consumer goods. There is also greater demand for materials to package these products. Many of these materials are limited and, at the same time, many of them pollute the environment.

9.3. Man and Renewable Resources

Most of the Earth's resources are not distributed equally. Air, however, is free and present everywhere. Air quality depends on freedom from pollution. The more industries a country has, the harder it is to keep air free from dust and chemicals. All living things need clean air to survive. Some air pollution comes from natural resources; some come from human activities.

Acid Rain. About twenty years ago, the Americans had a big project to repair the statue of Liberly. The metallic sheeting, which it was covered in, was slowly being eaten away, the main reason being the acidic rain in New York City. The acid rain dissolves marble and concrete, it accelerates the rate of rusting of sheet bridges and whatever else. Acid rain has some pretty devastating impacts on our natural environment as well. In some lakes the fish can no longer survive. Many forests can suffer from it. Some of Canada's most beautiful wilderness areas are showing effects of acid rain. American pollution is carried with the wind and blows into Canada where it falls as acid rain. The main source is the burning of fossil fuels. Coals, oil, and natural gas, being burned, give off a variety of pollutants, some of which are sulfur dioxide and nitrogen oxides. Both of these chemicals when combined with moisture in the presence of sun light form acids. Sulfur dioxide combines with water to form sulfuric acid. Nitrogen oxides react similarly to form nitric acid. These are the acids that are falling in the rain. They are not concentrated, but the levels are high enough to cause problems. Actually, unpolluted rain is slightly acidic due to natural sources such as volcanoes and natural non-metal oxides (carbon dioxide, nitrogen oxide, sulfur dioxide, etc.) occurring in the air. This is not a problem for the environment because over time it has adapted to this acid level. The changes that man has made have happened quickly, so that the environment cannot adjust that fast.

There are steps that can and have to be taken to solve the acid rain problem. One of the easiest and least expensive is the energy conservation. If we use less energy, less fuel needs to be burned. Simple things such as turning off lights, driving cars less, and recycling can make a big difference. Pollution control devices at the source is another way. Catalytic converters in cars reduce exhaust pollution. We should be willing to make sacrifices to ensure a clean environment for our own future and that of our children.

Greenhouse Effect (Global Warming).

There is growing evidence that the Earth's temperature is rising. This may be the result of the much talked-about "greenhouse effect". Scientists think that a greenhouse effect will cause frequent droughts in some areas and change dryer regions into crop producers. Flooding maybe another problem we have to face with the greenhouse effect. Slight increases in global temperature may melt polar ice caps and cause coastal areas to become covered with water as the level of the oceans rises.

The Earth's temperature is rising, because the levels of certain gases in the atmosphere are increasing. One of these greenhouse gases is carbon dioxide. In excessive amounts it traps heat. The high-energy ultraviolet radiation from the sun goes through the gases in the atmosphere. When it reaches Earth it is converted to heat energy. This is not as intense as the UV radiation, so it gets trapped by carbon dioxide when it would normally escape into space (Fig. 65).

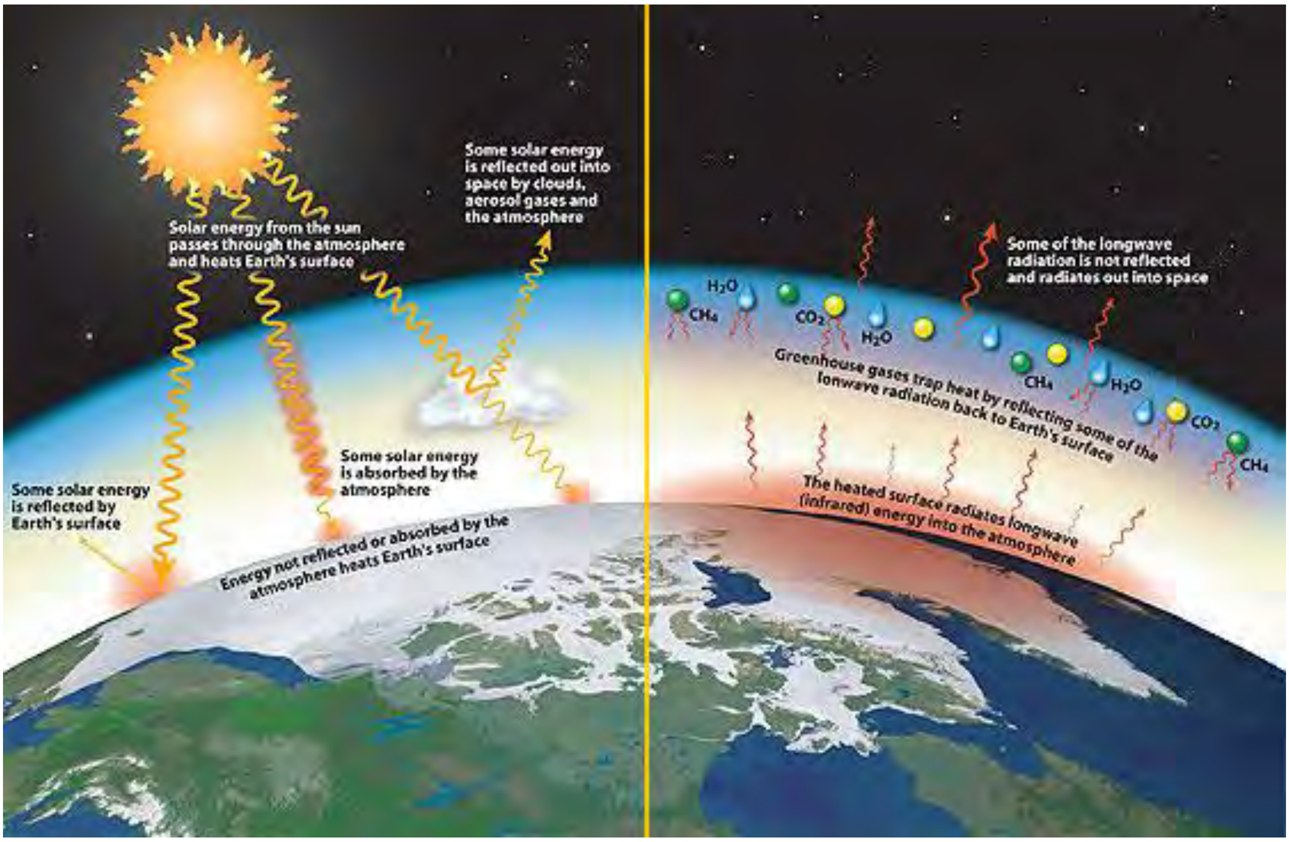


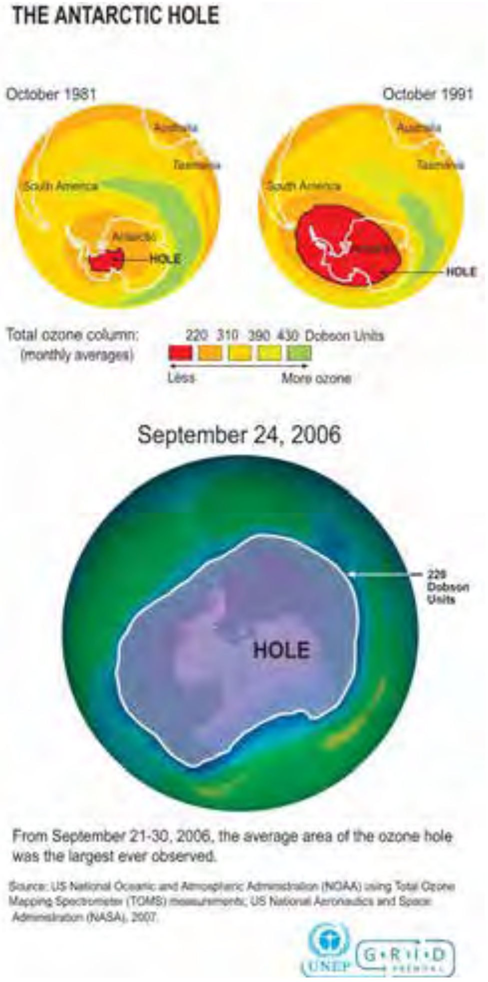
Fig. 65. How Carbon Dioxide Traps Heat.

Normally, carbon dioxide levels are kept in check by the plant life on Earth as they take it in during photosynthesis. But as forests are cut down to make room for housing and agriculture, we are decreasing the amount of carbon dioxide that plants can remove from the air. In addition, the slush and burn technique used to remove much of the world's rain forests releases more gas into the atmosphere.

The same conservation practices mentioned in the section on acid rain would also be effective in reducing global warming. As we conserve existing resources, we need to be looking at non-polluting alternatives. Hydroelectric power, nuclear power, and solar energy are some possibilities. They each have their drawbacks at this point in time, but if we are going to avoid the impact of the greenhouse effect we'll need to consider all the options.

The attention given to rainfall forests destruction may lead rain forests countries to be more careful with this precious resource. But these countries are poor and need help from other nations in order to survive without the income they receive from logging, or from the crops they plant once the forest has been chopped down. We are already seeing some evidence of the greenhouse effect, but the action taken now will determine its severity in the future.

Depletion of the Ozone Layer. There is a thin layer of gas around the Earth called the "ozone layer". It protects the Earth from excessive amounts of ultraviolet radiation. Without it life on Earth would literally burn up. Every flash of lightning converts some oxygen into ozone. Very high energy radiation from the sun strikes oxygen in the Earth's atmosphere and converts some of it to ozone. Despite this, scientists fear that the ozone layer is becoming depleted (Fig. 66).



From September 21-30, 2006, the average area of the ozone hole was the largest ever observed.

Source: US National Oceanic and Atmospheric Administration (NOAA) using Total Ozone Mapping Spectrometer (TOMS) measurements; US National Aeronautics and Space Administration (NASA), 2007.

Fig. 66. Ozone Layer in Relation to the Earth (Ultraviolet radiation is partly absorbed by the ozone layer).

The most obvious evidence of ozone depletion has occurred over the South Pole. In 1985, scientists discovered a hole about a size of the United States. This “hole” showed a 50% reduction in the amount of ozone. If it occurred anywhere else on Earth the effects would have been devastating. In fact, the winter of 1990 in Southern Australia brought a definite increase in skin cancer rate (It is summer in Australia when it’s winter in Ukraine.). Apparently the ozone hole in Antarctica has shifted over to the southern coast of Australia.

Certain chemicals, that we put into the air, damage ozone. They are called chlorofluorocarbons or CFC’s. CFC’s are used in refrigeration systems, foam plastics such as Styrofoam, cleaning solutions for electronics equipment, and aerosols. They constantly seep into the atmosphere and eventually end up high above in the ozone layer. Here chlorine from the CFC’s breaks down ozone molecules.

Air at ground level contains less than 1 part of ozone per million parts of air (1ppm). 24 kms above the ground there is a more concentrated layer of ozone. This layer contains about 6ppm of ozone. But it shields the Earth from much of the Sun’s ultraviolet light. Scientists use spectrographic analysis of satellite images to determine the amount of ozone in the atmosphere. Passenger balloons may travel in the stratosphere and air samples can be taken to measure the amount of ozone present. Weather balloons and high-level balloons sent up on a tether have also been used for the purpose.

After the hole over Antarctica was discovered, nations of the world moved quickly to determine how to handle the problem. A meeting in Helsinki in 1989 supported a total phase out of CFC’s by the year 2000. Still something must be done about the sources still in existence. McDonald’s no longer uses Styrofoam to package their hamburgers. Mass media continues to drown us in information about many aspects of air pollution. But “the cart still remains where it stopped”, as the Russians like to say (in original Russian it sounds: A voz I nynie tam).

Land Degradation. Worldwide, only 30 % of the land is inhabitable for humans while 11 % has the desirable climate, terrain (landforms) and soil necessary for prime farmland. The amount of land suitable for housing, farming, recreation and critical habitat for wildlife is a very small portion of what we actually see. Unfortunately, uses for suitable land often overlap. The areas good for farmlands are also prime housing area. In many cases these same areas are critical wildlife habitat. So we tend to treat it as goods to be bought and sold, or used as the highest dollar amount for the landowner. We need to ask ourselves if decisions based on economic values alone are wise ones. What about uses of land to meet human needs in the future?

The land is harmed not only by erosion but by things we put into it during waste disposal, the way we take thinks out of it through mining practices, and also by using land through agriculture. Erosion removes top soil and this makes land less fertile. Erosion may display itself in gullies and landslides. It is such a problem, that 2.3 billion tons of top soil are lost world wide per year. In our Ukraine it is 0.5 billion tons! In the USA, 20 % of original cropland is no longer in use due to erosion.

Agriculture is one of the main causes of erosion and also receives some of the impact. Overgrazing by livestock also results in severe erosion problem. Nothing is left to prevent topsoil from blowing away or flowing away with the rain water.

Deforestation is occurring at an alarming rate all over the world. Runoff from a forested slope is about 50 % less than that of a grass covered slope. As mountainous areas are deforested (as in our Carpathians), the water and soil that runs off can cause flooding and silt build-up in aquatic areas below. Forest fires are now many and very severe all over the world. Surface mining for the earth's resources causes erosion in a big way. Mines are a particular problem because much of the exposed soil contains metals and other toxic substances that can make their way into water supplies. Acid leaching from open-pit coal mines may also cause a lot of trouble. Erosion is also intensified by the building construction that goes along with urbanization.

The practices of erosion control are many and form a system. Such systems are studied in the course of Soil and Water Conservation.

Toxic Dumps. Many chemicals are dumped. The sites for that are seldom sufficiently secure. The containers leak and even though the area is capped with clay, some very dangerous stuff are released into the environment. The dissolved wastes get into basement walls, surface in backyards, and evaporate into the air. Over 300 chemicals were indentified is soil and ground water. Many are known as carcinogens.

Toxic chemical wastes come from a variety of sources. Industries that produce plastics, soap, synthetic rubber, fertilizers, synthetic fibers, medicines, detergents, cosmetics, paints, pesticides, and there is no end to this list. Modern people cannot do without industries, though we have a post-industrial epoch of our development. But we, the mankind, are to blame that each year we throw away or wash down the drain tons of toxic materials. There are three methods of land disposal for toxic waste currently used, deep well injection, surface impoundments, and landfills. New technology has helped to ensure that these methods will isolate wastes, and any leakage will be monitored and taken care of before damage is done. To reduce the actual amount of waste several things can be done. Companies can change the way they produce materials so that less wastes are produced. Waste materials can be destroyed using incinerators which burn up the waste, but this adds to air pollution. A better process called bioremediation has been developed. Special types of bacteria "eat" toxic wastes. We'll be seeing more of this type of clean up in the future as new microorganisms are developed to eat more and more types of waste.

Sanitary landfills are a good way out of the waste problem. About 80 % of USA trash is buried under thin layers of earth in sites known as landfills. They are quickly filled up.

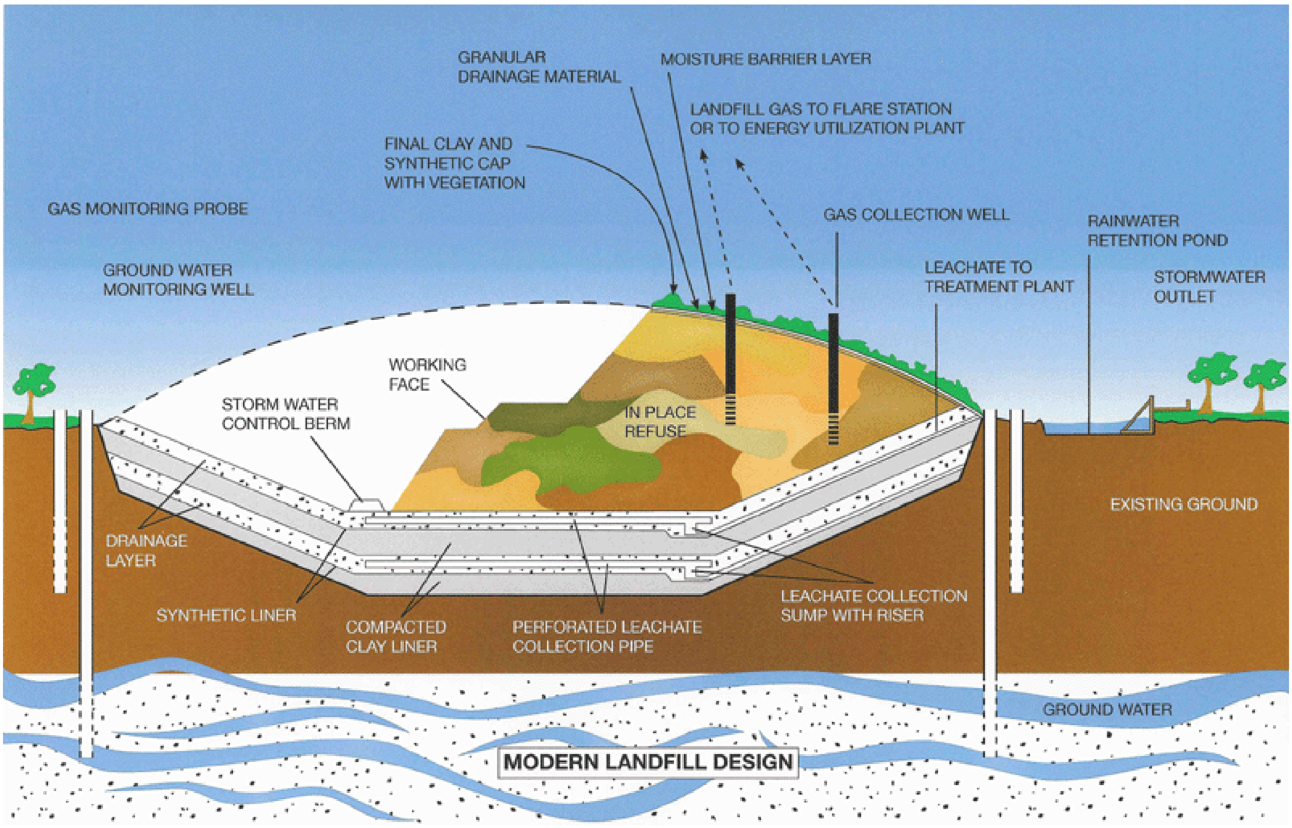


Fig. 67. A Sanitary Landfill (source: www.enviterre.com/services/).

As bacteria break down garbage, they give off methane gas which can seep through soil and into nearby houses. Settling is also a problem. But people are better at constructing landfills now than they used to be. Modern landfills are sited on high ground, above the water table. Gravel layers are used to vent methane, and sometimes that methane is actually pumped out and used to fuel. Garbage can be used to meet our energy requirements. Special plants are needed. They could incinerate garbage to produce electricity. But these are no problems of geology.

Water pollution. The demand for water increases even more rapidly than population increases. Dams have been constructed and reservoirs formed to provide a constant supply of water, but unfortunately, wastes produced by our growing population are rapidly polluting our water! Eutrophication, occurring naturally normally takes many years. But when man interferes, the process is accelerated and the natural balance is upset. Many types of pollution, beside sewage, can cause eutrophicaton, for example, the phosphates in fertilizers and detergents. The impact of phosphates became apparent when Lake Erie began to die. Algae began to overgrow, blocking out sunlight, and adding even more nutrients as they died and decayed. Once phosphates were banned however, life in the lake started to recover.

There are other types of water pollution. Toxic waste may leach out of storage sites on land and get into groundwater supplies. Some of the most serious of these are heavy metals such as mercury and lead, which got into our water after Chernobyl disaster. One type of water pollution that we hear about in the news is oil spills. An oil spill is a devastating thing for the environment. It poisons sea animals, destroys the fur and feathers of nearby land mammals and birds, and remains trapped in sand and sediments for many years. It is almost impossible to clean up. Thermal pollution (hot water), can be a major disruption to aquatic life. Electric power plants are a major source of the thermal pollution of natural water ways. The best idea is to prevent it all.

Eric P. Eckholm in his Losing Ground (1975) already stated that pollution problems "are but skirmishes compared to the uncontested, unpublicized routs being suffered in the hills of Nepal and Java, and on the rangelands of Chad and India. Both food production and economic development prospects in Africa, Asia, and Latin America are dimmed by accelerating destruction of the land's productivity". The author speaks about a tremendous deforestation, encroaching deserts, firewood crisis, salting and silting of irrigation systems and a crisis of world fisheries. Global food shortages, and hence inflation, will dump us into poverty and destitution if the undermining of food-producing systems is not arrested.

9.4. Man and Nonrenewable Resources

Nonrenewable resources include ores, oil, gas, coal, etc, the stores of which are very limited on earth. We must gain a proper understanding of the limitations of our nonrenewable natural resources. And, besides, so many mines are abandoned before all the ore is removed. What we call a mineral resource or ore depends on three things. First, does the deposit have enough of the valuable mineral to pay for the work of getting the ore to the market? Second, is the kind of mineral present, one that is in demand? Third, will the price paid for the mineral allow a profit? Some abandoned gold and silver mines have reopened. These mines are now profitable because the price of gold and silver has risen greatly.

But, as early as 1556, people were complaining about the effects of mining on their environment. Their concerns were similar to those heard today. Agricola (Georg Baner) wrote the earliest known book on mining called De Re Metallica. In this book he wrote: “But besides this… the fields are devasted by mining operations, for which reason formerly Italians were warned by law that no one should dig the earth for metals and so injure their very fertile fields, their vineyards, and their olive groves…” But how could the world’s population be fed without metal tools? The debate continues to this day. What are the current pollution problems of mining operations? First, mining displaces tons of rock. In the United States alone, over 13,000 square kilometers of surface area has been moved from above the mineral resources, the area being equal to the size of the state of Connecticut. Strip mines can and should be reclaimed, which means properly shaped and revegetated. Mined areas do not need to be eyesores. Where strip mines are properly handled, the soil layers are removed carefully and saved. After the ore is removed, the piles of waste rock are returned to the open pit. Then the topsoil is spread over the surface and replanted. In this way the area is restored to relative beauty. Some beautiful parks result from restoration of strip mines.

Some mining activities move only small amounts of rock with little surface damage. Quarries are open pit mines from which blocks of sandstone, linestone, or marble are removed. Clay, sand, and gravel pits usually are small pits that do not have any waste rock (кар’єри). The holes may become ponds used for recreation or they may fill with vegetation. The landscape appears to have a natural hollow or depression.

There may soon be a shortage of many materials now being buried as solid wastes. These include minerals such as lead, tin, chromium, nickel, manganege, etc. We must begin to reuse much of the waste we now discard. A more careful disposal of throw away items would help in two ways. First, the total volume of solid waste would be reduced. This would ease the problem of where to put waste without harming the environment or people’s health. Second, by careful separation of such items as paper, glass, and metal, we could reuse or recycle many of our natural nonrenewable resources that in the past have been buried and forgotten. This pertains to the return of glass bottles for reuse. In the United States about 400,000 metric tons of copper are recycled each year (20 % of the annual production). Nearly one third of their iron production comes from recycling old automobiles and scrap iron. In the former Soviet Union, young pioneers (scouts) helped a lot in this sphere of industry! Recycling lead, mercury, and aluminum accounts for lesser amounts of the yearly production of these metals. But some calamities are connected with this kind of activity in present day Ukraine, where poor people earn their scanty money on it. This is an activity for professionals, not bums and alcoholics. All this pertains to the flood of plastic bottles and old cans. One commandment the children should learn from parents and teachers is "Minimize your impact on the environment".

Modern recycling is a rapidly growing industry. It becomes even more important with every passing year. The future economy of the world depends on recycling for many material to make new products.

9.5. Man and Energy Resources

Energy is the ability or capacity to do work. Water power was one of the earliest energy sources. Today energy from running water is used to generate electricity. Most of the large rivers have dams build across them. At each dam, a plant uses the water to run generators which produce electricity. Nonetheless, this method of producing electricity meets only 10 % of our total energy demands. River flow regulation by the dams may accelerate the processes of silting. Large-scale construction for the purposes of water management and power generation, especially the regulation of river flow by the dams, reservoirs, and canals has brought to life some unheard of before by their magnitude and rapidity changes in the natural environment in the river basins. A great attention is deserved by the Report of the World Commission on Dams (2000), placed in the Internet which subjects to sharp criticism the negative consequences of dam construction for the environment. Soil desertification in river deltas is just one of these consequences.

Wind power has been used for centuries. About 70 years ago in the United States of America, almost every farm had a windmill to pump the water. In our Ukraine, the windmills were still in sight till the beginning of the sixties. Nowadays, as the cost of electricity increases, windmills are being used again. We can not expect to run our industries on wind power, but we have to resort to them all the same.

Solar energy constantly reaches the earth where it is used by plants to produce food. But most of the solar energy remains unused. We need to find some way to store energy for times when the sun is not shining. One method is called the solar cell. The cell collects energy from the sun and transforms it directly into electricity.

Nuclear energy. Nuclear fuels provide another source of energy. The fuel for fission power plants comes from uranium 235. Once the ore is mined, the uranium is processed into fuel rods or pellets. The fuel is loaded into the reactor core of the power plant. Uranium-235 nuclei break apart spontaneously. As the nuclei split, they form new elements and release vast amounts of heat energy. This energy is used to turn water to stream. Then the steam turns turbines, and the turbines run generators which produce electricity. The water is cooled and recycled. Nuclear power generation has problems of safety, waste storage, and fuel availability. For every 100 grams of uranium used, 99 grams remains waste. These waste products release radiation which destroys cells and changes or destroys genetic material. The waste must be stored until it is no longer harmful. This storage period ranges from 600 to 1,000 years or more. No one is sure how safe our storage methods are. Everyone is aware of the reality of Chernobyl disaster.

Fossil fuels include the hydrocarbons, coal, oil, and natural gas. Fossil fuels were formed during the decay of plants and animals that lived millions of years ago. It decay occurs where no oxygen is available, hydrocarbons are left behind. We now feel more and more, especially here, in Ukraine that fossil fuels are nonrenewable resources.

Coal, peat, lignite (brown coal), bituminous coal, anthracite, or hard coal, as well as petroleum and natural gas may not last forever. Some say that coal supplies should last for hundreds of years. As the supply of oil and natural gas decrease, coal is again becoming an important source of energy. We must find ways to reduce the amounts of sulfur, ash, and smoke which burning coal produces. Cities like Pittsburg, Pennsylvania, London, England, and St. Louis, Missouri, have had very smoke-blackened buildings and badly polluted air, which we have in Alcherske, because of the use of coal. Now they are cleaner, of course.

Petroleum or crude oil, and natural gas are preferred to coal because they are cleaner and easier to transport. Within the last 70 years, these two fuels have replaced coal for most home heating and transportation. No wonder our supplies are decreasing rapidly. Large amounts of crude oil are refined into petrochemicals.

Oil companies spend billions of dollars each year searching for deposits oil and natural gas. Most deposits are found between 500 and 6,000 meters below the surface. The search for oil covers large land areas and continues into water at least 2,000 meters deep. The scale of land and water pollution, connected with a mere search, is very large. There were times, when natural gas was burned at oil wells as waste. Now it is a very important fuel, as it is easily transported in pipelines. While these fuels still exist, we need to find replacement for them.

Future sources of oil. There is a large oil shale deposit in the Green River Basin of Colorado, Wyoming, and Utah. The process for removing petroleum from shale is costly. However, oil shales will be used in the future, even though they produce a tremendous amount of waste rock, which is an environmental problem. Tar sands form a large deposit in Canada. This oil is heavy and looks like the tar used in roofing. The deposit is so far north that the tar is hard to handle. It becomes stiff and hard in the cold climate. Geologists estimate that at least 300 billion barrels of oil can be recovered.

Coal can be transformed into gasoline, which in combination with alcohol forms gasohol. Cars can use this fuel with as much as 20 percent alcohol. Plant materials should be employed to produce alcohol, the waste products of plant materials.

Tidal power is a possible means of producing electricity and geothermal power is useful in some areas. Wood chips should not always be allowed to rot, though it is useful to remember that the co-called ramified chipped wood is a good fertilizer and a source of organic carbon for the soil. Be it as it may, it is profitable to think that solar energy may be the final answer to our energy needs and to try to think “decision-makingly” along this line is very important! The world’s supply of oil and natural gas may be gone very-very soon! Maybe in a couple of decades (in a score of years).

Questions and Assignments:

1. Explain why geologic activity of man may now become stronger than any of the natural geologic forces.
2. What can you say about the dynamics of world population growth in 19th and 20th centuries?
3. What can the scientists mean by “anthropogenic badland”?
4. Which natural resources are renewable and which nonrenewable?
5. Can you prove that the soil is also a nonrenewable resource?
6. Speak on the damage caused by acid rain.
7. What might be the consequences of global warming?
8. Speak on the depletion of ozone layer and its consequences.
9. Speak on the agents of land degradation.
10. What kinds of natural resources pollution do you know?
11. What should we do to economize our energy resources?
12. What is the final answer to our energy needs?

Part Two.

Soil Science

INTRODUCTION

Soil science is a science about soils, their origin, properties, composition, geography, development, functioning in the environment, rational use, conservation and reclamation.

It is well khown that the concept of soil as an independent natural body was introduced and developed by V.V.Dokuchaev (1846-1903) and his school of Russian soil scientists (A.P. Kostychev, M.M. Sibirtsev and others). Western Europe learned about it from the book D.K. Glinka published in German. Marbut's translation of this book made the Russian concept of soil as a natural body easily available to the English – speaking world. Recognition of the relationships between soil features and soil formation factors (climate, vegetation, parent material, landforms, and time) was developed not only by Russian scientists. Many other foreign scientists in the world recognised soil as a natural body (United Staties, Hilgard described the same things in his book had noted this asociation in a book published in 1860 (Ienny, 1961), and Coffey had recognized by 1912 (Kelloy, 1941). However, the development of the concept of soil as a natural body with horizons that reflect the influence of soil forming factors, particulary climate and vegetation, may be credited to the Russian school. Ukrainian Soil Science is traditionally genetic as it is based on the study of origin, development, and evolution of soils developed by the scientists from the former Soviet Union.

Soil and its definitions

H.D. Foth in his Fundamentals of Soil Science states that soil can be also viewed as a mixture of mineral and organic particles of varying size and composition in regard with plants growth. Today over half of the world's population till the land. The concept of soil as a medium of plants growth was born in ancient times and still remains "valid" to this day. But even the farmers understand soil as a natural body and the object of labour (tillage fertilizing, reclamation, etc). It is needed to be understood form the very begining that in any soil formed in a mineral parent material the solid particles constitute about 50% of the soil's volume. The remaining soil volume (about 50%) is represented by porous space composed of pores of diferent shapes and sizes. The porous space contain air and water and serve as channels for their movement as well as for the growth of roots. Roots anchored in soil support plants and absorb water and nutrients. The three essential things that plants absorb from the soil and use them are as follows: (1) water that is mainly evaporated from plant leaves, (2) nutriens for nutrition, and (3) oxygen for root respiration. The ability of the soil to produce yields of crops (its fertility and productivity) depends on the following factors of soil fertility graphically depicted by the Ukrainian soil scientist O.N. Sokolovsky (Fig. 68).

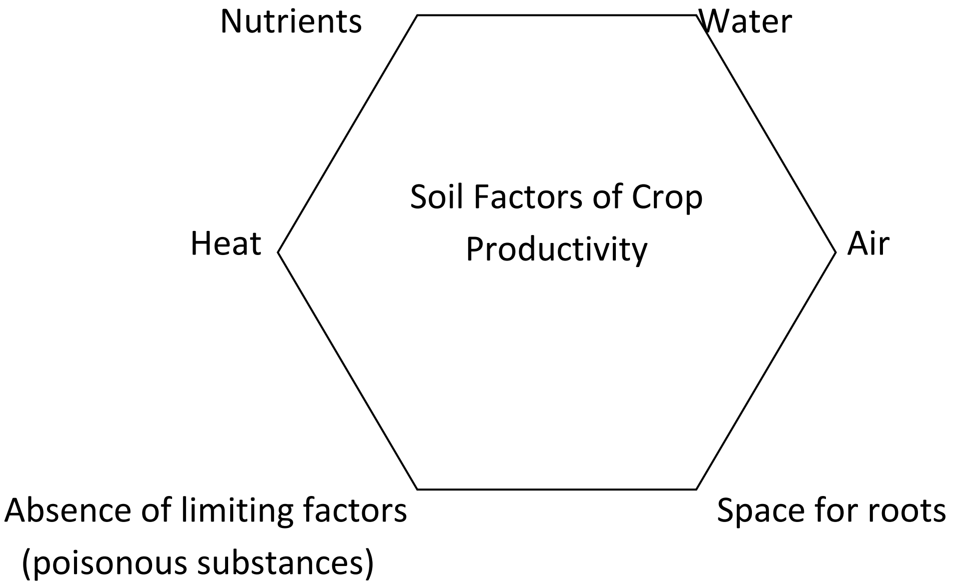


Fig 68. O.N. Sokolovsky’s Interpretation of Soil Factors of Crop Productivity (1986)

In his last fumdamental work, *Lectures on Soil Science* (1901), V.V.Dokuchaev wrote that the soil is “the fuction (result) of parent material (subsoil), climate, and organisms multiplied by time.” According to V.V. Dokuchaev, the soil is not only an independent natural body, different from all the other natural bodies, but also an historic phenomenon having its age and the history of origin and development.

V.V. Dokuchaev was very constructive in identifying the existence of functional relationship between the soil and all other natural bodies and phenomena. B.G. Rozanov, a noted Russian pedologist, gave the most exhaustive definition of the term “soil” according to which “the soil is **endowed with fertility, complete polyfunctional, and policomponent open multiphasial structural system in the surface layer of the cortex of weathering, as well as a complex function of parent material, organisms, climate, relief of landforms, and time”.**

Soil as a Specific Natural Body

Soil occupies an intermediatate position between the living and nonliving natural bodies.V.I. Vernadsky (1863-1945) attributed it to the specific biomineral, or biotic – abiotic natural bodies. Some scientists think that the soil itself is dead. But it is not quite true, as the soil is the home of innumerable forms of plant, animal, and microbial life. According to P. Farb, “we live on the roof tops of a hidden world”. Life in the soil is amazingly diverse, ranging from microscopic single-celled organisms to large burrowing animals. Between all of them there are certain food chains and competition for survival. The multiphasial nature of soil is reflected in the existence of four phases: (1) solid, (2) liquid, (3) gaseous,and (4) living or “alive”.

Soil Individuum (Pedon)

Soil is a natural body with a certain extension in space in three dimensions. As any natural body, it has its location in space and volume boundaries. Soil body occurs as a continuum on the surface of the earth. Many scientists define soil as “naturally occuring, unconsolidated mineral or organic material at least 10 cm thick that occurs on the earth’ soil surface and is capable to support plant growth” (Canadian System of Soil Classification, 1987). According to P.S. Kossovich, the lower boundary of the soil is determined by the depth to which the initial rock underwent transformed during in the course of soil formation. The upper boundary of the soil is the surface dividing the soil from the atmosphere for the terrestrial soil or the surfare between the soil and the hydrosphere for the submerged, marshy, peddy rice, and other soils. There are lateral boundaries between the soil units which are usually rather diffuse, never the less a diffuse boundary is also a boundary.

The pedon, as defined by pedologists of the United States (Soil Survey Staff, 1975), serves as the basic unit for soil classification in many countries. It is the smallest three dimensional unit on the earth surface that is considered as soil. Its lateral dimensions differ, but the pedon is the minimal volume of the soil the horisontal dimentions of which are sufficiently large to have a complete spectre of variability of the ratios between the horizons corresponding to the minimum horizontal diversity of the soil by the diagnostic features (Fig. 69).

A contiguous group of similar pedons is called a polypedon, which is included in soil taxonomy (Soil Survey Staff, 1975) as a unit of soil classification. All of the variability within the pedon is included in the concept of the taxonomic class used from order to series. Russian pedologist F.I. Kozlovsky (1972) measured the areas of soil pedons (in m2) for the soddy-podzolic soil in the fluvial loam (Spodosol) of Podmoskovye to be within 11.75 to 28, whereas for the typical chernozem (Mollisol) of Kursk region it was about 28 and for meadow – chestnut soil of Sarpinsk Lowland, formed in “chocolate clay” parent material, it fluctuated around 7.

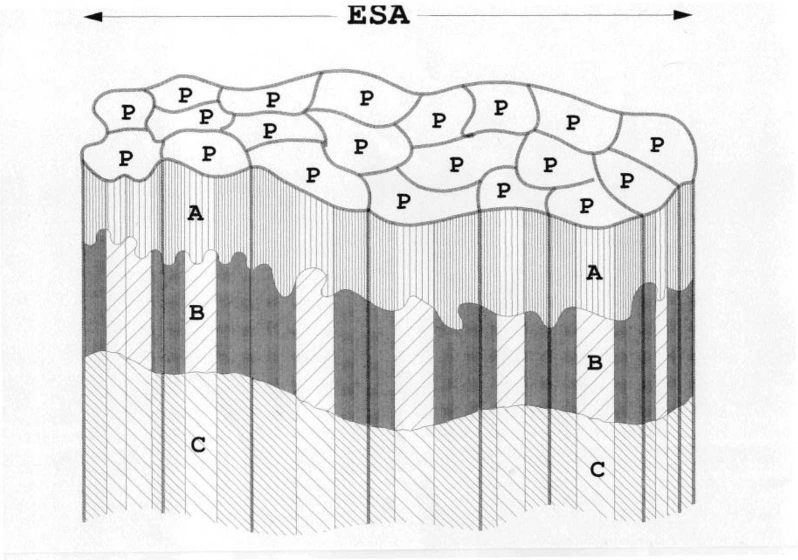


Fig. 69. The position of pedons in an elementary soil areas areal (ESAA) or polypedon. Elementary soil areals are the mapping units in a large scale soil mapping in Ukraine.

Place and Role of Soil in the Environment

Soil cover (pedosphere) lies on the boundary of such planetary spheres as litosphere, atmosphere, and hydrosphere and develops itself in close relationships with them. Soil cover is a component part of the biosphere (ecosphere) which is the sphere of the life forms. The figure below illustrates the place of soil cover (pedosphere) in the environment.

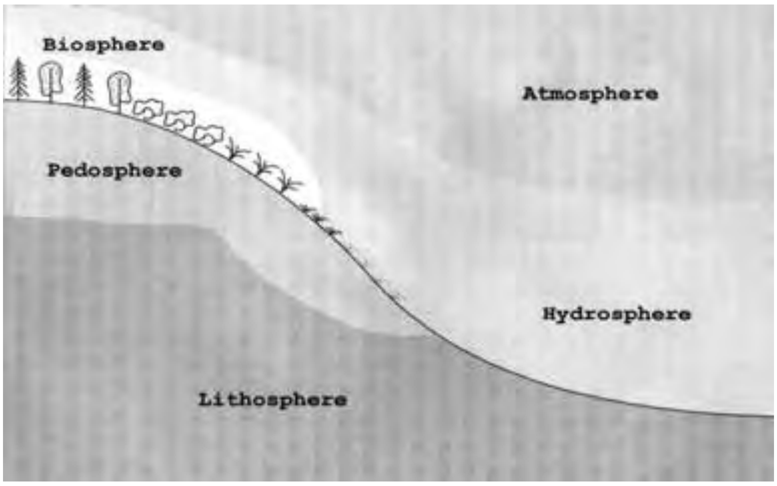


Fig. 70. Pedosphere in the system of other geospheres

The soil cover of the Globe performs several the so-called global functions.

The first and essential of these functions is to sustain life on the Earth. Plants absorb the nutrient elements from the soil, wereas the animals and human consume them with plant products. Soil accumulates the biogenic elements needed by the organisms in the available forms. Terrestrial plants take root in the soil. Higher plants are the major producers contributing to the supply of soil organic matter (SOM). The microorganisms (bacteria and fungi) are the major decomposers and mainly responsible for the cycling of nutrients and energy in soil ecosystems. Soil animals are also numerous and diverse. Earthworms, ants, termites and rodents move large quantities of soil and may greatly alter the nature of soil horizons. Soil is the consequence of life and at the same time a primary condition for the existence of life on the Earth.

The second global function of soil is to make possible the continuous interaction between the great geologic and small biologic cycles of substence and energy on the surface of the Earth. Getting to the surface the primary (igneous) rocks undergo weathering and accumulate the nutrient elements for the living organisms. The elements are absorbed by the plants and through the series of intermediate trophic cycles (plants → animals → microorganisms) return back to the soil which is the essence of the small biologic cycle of the substances. The elements are partly leached from the soil by atmospheric precipitation into the stream pattern and the zones of accumulation. Finally they reach the world ocean where they may give rise to the formation of sedementary rocks which, in their turn, may reach the terrestrial suface or subjected to the metamorphism. That is the great geologic cycle of substances or rock cycle. Soil perfoms the function of a linking chain and a regulator of the interaction between the two cycles of substances on the surface of the Earth.

The third global function of the soil is to regulate the chemical composition of atmosphere and hydrosphere. Soil respiration complimented by the photosynthesis and respiration of living organisms plays a determining role in the maintaining of the atmospheric air composition. In the Earth’s geological history the soil played an important part in the creation of the existing atmosphere. Soil cover also affects the composition of substances which reach the hydrosphere on the continental branch of the global water cycle.

The fourth global function of the soil is to regulate the life sustaining processes in the biosphere. In this respect, the geographic distribution of soils regulates the “density of life” on Earth. The dynamic reproduction of soil fertility is very important for such regulation. The propertities of soil may sustain the life of plants and to limit it. The spread of terrestrial life is determined by the climate and modified by the properties of soils.

Finally, the fifth global function of the soil is to accumulate an active organic matter and chemical energy retained in it. According to V.N.Sukachev (1880-1968), the soil is an inalienable part of natural ecosystems or **biogeocaenoses.** Almost all life in the soil is dependent on organic matter for nutrients and energy. The organic matter in an ecosystem consists of the organic matter above and below the soil surface. It is mainly the production of the primary producers – the higher plants. Microorganisms play important environmental roles decompositing and supplying them with the nutrients.

Place and Role of Soils in Life and Human Activity

People get 99% of their food and fiber from the soil. For the last centuries the world population growth exceeded by its kinetics the natural growth of biologic populations described by the simple first order differential equation: dP/dt = KP, where P = population number and t = time. The law of the world population growth could be described by the equation: dP/dt = KPt and since approximately 1827 to 2000 the well population of the globe could be roughly computed by the partial solution of the above stated equation which is the following formula:

$$P_t = \exp\left[6 \times 10^{-5}(T-1827)^2\right] \qquad (12)$$

where *T* is the year of enquiry within 1827-2000. Calculated by this formula, the world population in 2000 should have been 6 billion people, (and so it had been), whereas in 1827 it was only 1 and in 1950 – only 2.5 billion.

In the last decade the curve of the world population growth passed through its inflexion point and the rate of the future growth will become increasingly smaller. The increasing demands of humans for food and fiber make it more and more difficult for the soil to meet these demands. The pollution problem also has to be remembered. The rapidly growing world population is polluting and squandering the resources that support life, and in this respect the soil is better to be considered a non-renewable resource.

The soil still remains the basic and unreplaceable means of agricultural production. E.P. Eckholm (1975), speaking on the environmental stress and world food prospects, states: “Coaxing enough food from the earth has traditionally been guided by a certain simple logic: plow more land, intensify labor, refine techniques, and the supply of food will grow commensurately. But this has been the logic of humans, not of nature.” The past few decades have brought an end to the traditional expansion onto new lands as a major means of increasing food production. It appears that an expanding world population will have to produce its food on a declining amount of land per person. Land requires considereable investments to remain productive, and is subject to degradation by erosion, pollution, salinity, waterlogging, and desertification. H.G. Foth in his *Fundamentals of Soil Science* states, ”… the main limitations for using the world soil resources for agricultural production are drought (28 %), minerals stress or infertility (23 %), excess water (10 %), and permafrost (6 %). Only 11 % of the world's soils are without serious limitations”. The settlement of the problem of food shortage and environmental stress can hardly be considered as optimistic.

The urgent task is to provide the economic, social, and political environment for the solution of this problem.

History of Soil Science

A famous pedologist I.A.Krupenikov in his unique publication *Soil Science History* (1981) identifies the follows main periods in this science history:

(1) **Period of the initial accumulation of seemingly unrelated facts on the properties of soils, their fertility, and the ways of management.** This period was connected with the origin and gradual development of land farming about 10-11 thousand B.C., when human beings were hunters and gatherers. It is impossible to determine where crops were first cultivated. In 1946 archeologists discovered an ancient village at Jarmo in northern Iraq (Braidwood and Howe, 1960). Stone implements found at the village site could have been used for tilling the soil and for weeding growing crops. This village was occupied about 11,000 B.C. and is considered the earliest site of neolithic cultivated agriculture. The first “farmers” could differentiate the land plots by their productivity and invent some primitive implements for soil tillage and other operations.

(2) **Period of Discriminative accumulation of knowledge on soils, their registration, and creation of the first primitive land cadastres.** The period lasted for several thousand years and coincided with the development of slavery.

The valley and the delta of the Nile in Ancient Egypt were the regions of favourable farming. The flooded areas in the fertile valley had a sufficient amount of moisture and a thick layer of black silt. Thousands of people living in nearby localities worked together building and repairing the dams and canals. Large areas in the valley and the Delta of the Nile were suitable for farming. The Egyptians sowed barley, wheat, and flax; they also grew vegetables. The greatest agrarian and political achievement of the period was determining the origin of land evaluation for the purpose of differentiated taxing. Some papyruses were preserved and came to our days with the descriptions of land and weather in certain localities of Egypt.

The territory between the Euphrates and the Tigris, in their middle and lower reaches has always attracted attention of soil scientists. The fertile clay soil of Southern Mesopotamia attracted farmers. The first settlers appeared several thousand years B.C. The farmers dug canals to irrigate their fields and dry up the swamps. They surrounded the settlements and orchards with dams to protect them from floods. They learnt to make copper and then bronze work tools and invented a plow, which could be used on heavy, clay soil. Landowners appeared which realized the advantage of having slaves. By the beginning of the 2nd millennium B.C., the powerful Kingdom of Babylonia was formed, which reached the height of its power in the first half of the 18th century B.C., in the reign of king Hammurapi. A black stone slab discovered by the archeologists early in the 20th century was covered with inscriptions on both sides. They were the **laws** issued by king Hammurapi and among them were the laws of land and water ownership, proper management, and protection. Some clay slabs were extracted from the Babylonian earth with the maps and drawings of irrigation systems.

(3) **Period of primary systematization of the knowledge of soils connected with the civilization of ancient Greece and Rome.** The period lasted for a millennium since the 8th century B.C. Ancient Greece and Rome were known for a rapid development of science and arts. Knowledge on soils, their management, practices of crop production, viticulture and etc had been generalized in the form of philosophical and poetic literary works.

The first soil classifications were based on the soil use in farming. Soil descriptions of the various regions and the ways of crop production found their reflection in the works of Theophrastus and Eratosphene. Columella wrote a philosophical work entitled *On farming* (*De re rustica*), consisting of four books, which proved to be a veritable agronomical encyclopedia. Mark Portium Caton was the author of the “De agriculturae” which generalized the experience of any generations of free farmers.

Under the slave-owning system, a huge labor force was placed at the disposal of the owners. In spite of the low technical standards, people excavated huge canals and erected dams, transformed swamps and deserts into flowering orchards and fertile fields, worked mines and quarries, and built large ships. Large scale irrigation was indispensable in the East, as well as protection from the floods. Only large and powerful states could procure hundreds of thousands of peasants and slaves and force them to work for the benefit of a small class of slave-owners. Rome developed into a slave-owners' and aristocratic republic. It was impossible under the slave-owning system to improve the methods of land cultivation, or to introduce new inventions. A better plow was invented in Rome, which made land cultivation more productive. This plow was mostly used in provinces inhabited by free peasants. Slaves were given nothing but old, heavy plows. Poor cultivation exhausted the fields. Harvests deteriorated from year to year.

(4) **Period of intensive land cadastre activities lasted during 15-17 centuries of the new era (A.D.) was connected with the development of land survey and soil evaluation for the purpose of feudal taxing.** The period was not as significant for the development of soil science as the previous period. Official land cadastre was introduced in the majority of feudal countries in a due time within the period. Albertus the Great (Germany) and Petras Crestentius (Italy) wrote their great books on soils and agronomy in the 13th century. They revived Columella's “know-how” and practical recommendations. Francis Backon in England wrote a discourse on the “water nutrition” of plants, whereas Bernar Palicci in France discussed the absorption of salts by the roots of plants from the soil. The great and versatile Leonardo (da Vinci) was the first to reasonably speak about the cycle of substances in Nature.

(5) **Period of intensive experimental and geographic studies of soils and their fertility** was connected with the rapid development of extensive farming and crop production on large areas of land. Of great demand in the 18th century was the book by J.A. Kuelbel *Cause of the fertility of lands* (1741), in which the author bad gave reasons on the main role of water in plant nutrition. Economic evaluation of land found its place in the work of French physiocrates. A.Turgo put forth the theory of diminishing returns (1766), severely criticized by the Marxist economists. Valerius in Sweden investigatiated soil humus and put forth the hypothesis or the essential role of soil organic water (humus) in plant nutrition. Ahard in Germany was the first scientist who fractionated soil humus to evaluate its quality. M.V. Lomonosov, a Russian academician, advanced some novel ideas on soil origin and development (1763) and in this respect he was promptly followed by P.S.Pallas (1773) also a Russian scientist, who put forth the hypothesis of chernozemic soil formation.

(6) **Period, in which Agrogeology and Agriculturchemistry** reigned supreme, coincided with an accelerated development of the capitalist mode of production in

the agriculture of Europe in the 19th century. This period preceedied the creation of pedology as a science. The founders of agricultural chemistry H.Davy (*Elements of Agricultural Chemistry*, 1813), E. Wollny *Discourse on Organic Matter and Humus formation,* 1892), A. Thaer, M.G. Pavlov, J. Liebig (*Chemistry in Application to Agriculture and Plant Physiology,* 1840) formulated the basic principles of this branch of science. Ch. Darwin in his *Earthworms and vegetable mold* (1881) demonstrated a clear understanding of the earthworms' role in soil formation. C Sprengel's wrote his *Die Bodenkunde* (1837), which was the first textbook of Soil Science and coined the term Bodenkunde (Soil Science). Another agrogeologist, F.A. Fallou, devoted all this life to pedology, carrying out comparative studies of soils and their parent materials, developed soil classification based on nature of parent material and exogenic processes. W.Knop (1817-1891) proposed his own soil classification based on its chemical composition. His three classes of soils were (1) silicatic, (2) carbonatic, and (3) sulfatic (gypsiferous). N.I. Zheleznow in Russia studied some aspects of soil mechanics. The first soil map in Russia was compiled in 1851 under the auspices of K.S. Veselovsky and the second one – in 1879, by the guidance of V.I. Chaslavsky.

(7) **Period of the modern genetic soil science's inception and first stage of its development** at the end of the 19th and the beginning of the 20th century was a **crucial period in the development of science.** Much is worth to be said about it. V.V. Dokuchaev proved the soil to be a unique body of nature, developing "historically" from geologic formations (parent materials) under the combined influence of water, air, and the organisms. As any other natural body, the soil has its constitution and structure, its specific properties and morphologic features, undergoes the stages of its specific life, has its age and low-governed distribution over the surface of the Earth.

V.V. Dokuchaev (1846-1903) formulated the basic, cornerstone principles of the up-to-date soil science: of soils as "self-sustained" natural bodies, like minerals, rocks or plants; of factors and conditions of soil formation, of zonality of soil cover (latitudinal and vertical), on the specific methods of soil survey, and on the fundamentals of soil mapping. The scientist lived a comparatively shorf life but was happy to see the proofs of his theoretical concepts in real life and in the practices of farming and land amelioration. His fundamental works were *The Russian Chernozem* (1883), *Our Steppes Earlier and Now* (1892), *To the Teaching on the Zones of Nature* (1899), *Materials on the Survey of Russian Soils* (1886), etc. V.V. Dokuchaev's student, M.M. Sibirtsev, and his constant opponent P.A. Kostychev deserve to be named as “cofounders of genetic soil science” in Russia. Much was done in other countries. E.W. Hilgard (1833-1916) and M. Whitney (1860-1927) of the USA were close to V.V. Dokuchaev’s concepts, though their standings were more “agrogeological”.

One of E.W. Hilgard’s important works was *Soils. Their Formation, Properties, Composition, and Relation to Climate and Plant Growth in the Humid and Arid Regions* (1906). M. Whitney wrote *Soils of the United States* (1909). European development of soil science in the period was connected with the names of E.Wollny, E. Ramann (*Bodenkunde*, 1911), A.Sigmond, B.Murgoci, W.Murchison (Dokuchaev’s predecessor, who wrote *On the Tchornoi Zem, or Black Earth of the Central Regions of Russia* in as early as 1842) and numerous other scientists. Academican K.D. Glinka published his profound and world-famous textbook of *Soil Science* in 1908.

(8) **Period of development and final establishment of V.V. Dokuchaev’s genetic soil science in the world** lasted between the two world wars (1914–1941). A large amount of information had been collected in various countries on chemical, physical, and mineralogical properties of different soils. The basic concepts of soil chemistry, physics, and biology were formulated into soil science was differentiated into a series of a sciences. Soil mapping developed at an accelerated pace. International Soil Science Society was founded in 1924 and three congresses (1927, 1927, and 1935) took place in Washington, Leningrad, and Oxford respectively. Rusian school of soil science still remained the leading in the world. Ukrainian soil scientist O.N. Sokolowskiy started his sceintific activities in that period too. But Ukrainian soil science developed under the influence and “theoretical guidance” of a gifted coterie of Soviet scientists, including W.R. Williams, K.K. Gedroiz, B.B. Polynov, M.I. Dimo, D.M. Pryanishnikov, E.N. Ivanova, G.N. Vysotsky I.M. Antipov-Karataev, etc. G.I. Makhov, a Ukrainian soil scientist, published the “Soils of Ukraine” in 1932.

(9) **Period of intensive development of soil surveys and inventory on the global scale** was characterized by broad international cooperation of researchers works and lasted for 30 years after the Second World War. The main feature of the period was an intensive survey of soils in the former colonial and semi-colonial countries. Under the UN organization systems (FAO, UNESCO) in the framework of activties, from 1960 till 1978. The soil map of the world had been compiled under the scale of 1:5000000. Dozens of international conferences and coordinating shopworks were held for discussing most challenging problems of soil science and pertaining to soil science, burning issues of environmental protection, etc. International Soil Museum

was created in Amsterdam exhibitity a rich collection of the etalon soil specimen. The translation of a vast volume of literature into all languages on all aspects of soil science was important in the development of new ideas of soil genesis and geography, their classification and taxonomy, and many new soil types have been described.

(10) **Period of intensive scientific activities in soil conservation and rational use** (an important constructive period in the history of soil science) started in the mid-seventies of the 20th century and lasts to the present time. The period is characterized by an ever wider understanding of the global scale of environmental problems. The soil cover of our planet is suffering from numerous degradations and rapid diminishing of areas good for agriculture. International soil research projects under the auspices of the UN organization systems are becoming ever more effective in fighting against desertification and hunger in the world. The rapid development of information (and geoinformation) technologies gives possibilities for the researchers to carry out a soil and enviromental surveys.

Present-day Ukrainian soil scientists still have to react to the challenges of our time and create the generally accepted and sufficiently effective for practice soil classification and taxonomy well harmonized with more widely spread counterparts in the world. No less great challenges are now faced by the soil and land quality evaluation in Ukraine, despite the great efforts taken in this field by V.V. Medvedev, M.I. Polupan and other prominent soil scientists.

SOIL COMPOSITION AND PROPERTIES

Chapter 10

Rock and Mineral Weathering. Parent Materials of Soils

10.1. Weathering

Weathering is a set of complex and diverse processes of quantitative and qualitative changes of geologic rocks and minerals composing them under the influence of atmosphere, hydrosphere, and biosphere.

Surface horizons of rocks subject to weathering are called the cortex of weathering. Two zones are identified within the cortex of weathering: the surficial or recent zone of weathering and deep or ancient zone of weathering. The thickness of recent cortex of weathering, in which pedogenic (soil-forming) processes might occur, may be in the range from a few centimeters to several meters. By the physical, chemical and biological factors involved, three forms of weathering were identified: physical, chemical, and biological, though biological weathering by its essence may be both physical and chemical.

Physical weathering takes place when a rock is reduced to smaller fragments without undergoing a change in chemical composition. This type of weathering, known also as disintegration, may be the result of a variety of physical forces. Frost action, alternate heating and cooling, freezing of water in cracks – all are the most potent agents of physical weathering. The most intensive physical weathering occurs under the influence of great gradients of diurnal and seasonal temperatures. In hot deserts the rock surface may reach the temperature of 60-70ºC during the day and cool to nearly 0ºC at night. Physical weathering becomes more intensive when water gets into cracks and fissures of rocks and creates capillary tension there. Freezing water increases its volume by one-tenth which produces a greet pressure upon the walls of cracks.

In arid regions the same action is caused by soluble salts, which crystallize from their solutions in cracks, fissures and other voids in the rocks. Anhydrite $(CaSO_4)$ combining with water molecules, transforms it into gypsum $(CaSO_4 \times 2H_2O)$ and increases its initial volume by one-third (33%). As rocks on the Earth’s surface undergo weathering, large pieces often break off the rocks. Over a long period of time, the rocks crumble and decay. One can see the results of weathering at the base of a mountain or on a mountain slope. Pieces of broken rocks pile up in these areas. These piles of rock fragments are called talus slopes. During mechanical (physical) weathering, rocks are broken into different shapes and smaller pieces. At the beginning of the weathering process, typical rock fragments are sharp and angular. As weathering continues, they become smooth and rounded. Often the pieces break off in curved sheets or slabs parallel to the rock's surface, which is called exfoliation. Wind-blown sand causes mechanical weathering of rocks by abrasion. Abrasion is the wearing away of rocks by solid particles carried by wind, water, or other sources. In some areas of the Earth, particularly in certain mountain regions, the rocks are subjected to drastic temperature changes almost daily. Rocks of high mountain peaks expand as they are heated in the daytime, and contract when subjected to freezing temperatures at night. In rocks with well-formed distinct crystals of various minerals possessing varying coefficients of volumetric and linear expansion, the temperature variations cause the disintegration of a rock into the grains of the minerals composing it.

As a result of physical weathering, solid rock becomes capable to retain and conduct through itself water and air. Its surface area increases considerably, which, in its turn, creates favorable conditions for chemical weathering.

Chemical weathering, or decomposition, produces a chemical breakdown of the rock, which may destroy the original minerals and produce new ones. Physical weathering simply produces smaller fragments of the initial (parent) rock. Chemical weathering produces rock materials that are basically different from the original rock. Although chemical changes occur in a variety of ways, the more common processes of decomposition are dissolution, hydration, hydrolysis, carbonation, etc. As chemical changes take place, minerals can be added to or removed from rocks. Many substances react chemically with rocks to break them down. Most chemical weathering is caused by water and carbon dioxide. Water can dissolve some of the minerals that hold rocks together. Rocks that dissolve in water are said to be soluble. Water can also form acids when it mixes with certain gases in the atmosphere. Carbon dioxide (CO_2), which is generally present in air, water, and soils, commonly unites chemically with certain rock minerals, greatly altering their composition. Substances produced in this manner (carbonates and bicarbonates) are relatively soluble and therefore easily removed and carried away. In addition, the union of carbon dioxide and water produces carbonic acid (H_2CO_3), an effective agent in attacking such minerals as calcite and dolomite:

$$CaCO_3 + H_2O + CO_2 = Ca(HCO_3)_2$$

In nature carbonic acid is formed when carbon dioxide in the air dissolves in rain. This slightly acidic rain falls to the ground and is able to dissolve certain rocks on or beneath the surface of the Earth. Sulfur oxides are a byoproduct of the burning of coal. These compounds dissolve in rainwater to form sulfuric acid. Rain that contains sulfuric acid is one type of acid rain. If corrodes or wears away rocks and other materials very quickly. The solubility of minerals in salt water also increases. Every increase in temperature by 10ºC hastens the rate of chemical reactions 2-2.5 times. This explains a very intensive chemical weathering in equatorial areas of the globe and their slow passage in polar ones.

The reaction of feldspar with water may illustrate the hydrolysis:

$$KAlSi_3O_8 + H_2O = KOH + HAlSi_3O_8$$

An acid environment stimulates the reaction of water with minerals and is one of the most important weathering reactions:

$$4NaAlSi_3O_8 + 18H_2O + 4H^+ = Al_4(OH)_8[Si_4O_{10}] + 8H_4SiO_4 + 4Na^+$$ Albite

kaolinite hydrosilica

In the reaction a primary mineral, albite, is converted to kaolin, a clay mineral. Some clay minerals tend to be resistant to further weathering. A process described above is sometimes called **kaolinization.** Chemical weathering of the most abundant primary minerals (orthoclase, albite, etc.) gives birth to a number of secondary minerals: salts, oxides (hydroxides), and clay minerals. The kaolinite, even though very resistant to further weathering, may be decomposed and disappear from the rock or soil:

$$Al_4(OH)_8[Si_4O_{10}] + 10H_2O = 4Al(OH)_3 + H_4SiO_4$$

Hydration of minerals (their chemical union with water) produces hydrons oxides:

$$2Fe_2O_3 + 3H_2O = 2Fe_2O_3 \cdot 3H_2O$$

An example of hydration may be the hydration of anhydrite into gypsum:

$$CaSO_4 + 2H_2O = CaSO_4 \cdot 2H_2O$$

The hydration of anhydrite increases its volume and solubility in water at 20°C from practically zero to 2.6 g•L^{-1}.

Oxidation occurs as oxygen, assisted by moist air, combines with minerals to form oxides. Rocks and minerals containing iron compounds are especially susceptible to this type of decomposition. The oxidation of iron compounds, which produces rust, is responsible for the color of many red, yellow, and brown rocks and soils. Certain iron compounds (for example, pyrite) form acids when oxidized:

$$2FeS_2 + 2O_2 = 2FeSO_4 + H_2SO_4$$

$$12FeSO_4 + 6H_2O + 3O_2 = 4Fe(SO_4) + Fe(OH)_3$$

$$2Fe_2(SO_4) + 9H_2O = 2Fe_2O_3 \cdot 3H_2O + 6H_2SO_4$$

The acid affects the rocks and thereby hastens the process of decomposition.

The joint action of oxidation, hydration, and carbonation is a powerful agent of weathering of amphiboles and pyroxenes:

$$\underset{Augite}{4CaFeSi_2O_6} + O_2 + 4H_2CO_3 + 6H_2O = 4CaCO_3 + \underset{goethite}{4FeOOH} + \underset{hydroussilica}{8H_2SiO_3}$$

Reduction processes also play a significant role in the weathering of minerals containing reducible elements in an oxidized form. But this process involves the participation of chemotropic microorganisms and belongs to biol+ogical weathering. The weathering of certain silicate minerals is very important for the formation of parent rocks and soils. Thus, ferromagnesian silicate weathering may be represented by the generalized scheme:

$$\begin{matrix} Olivine \\ Pyroxenes \\ Amphiboles \end{matrix} \xrightarrow{+O_2} \underset{(precipitated)^+}{Fe(OH)_3} + Mg^{2+} + Ca^{2+} + H^+ + silica \quad residue$$

Soils dominated by these minerals in the fine silt and clay fractions are the minimally weathering soils all over the world, but are mainly soils of the desert regions where limited water keeps chemical weathering to minimum. Minimally weathered soils are generally fertile. Muscovite mica weathering proceeds in the following way:

$$Mus\,cov\,ite \xrightarrow{-K^+} Illite \xrightarrow{-K^+} Vermiculite \rightarrow Smectite$$

Potassium cation released from micas and hydromicas (illite) becomes an available form for plant nutrition.

As to **organic or biological weathering**, some geologists are still undecided as to identifying it among the types of weathering, as its mechanisms, in their essence, belong to either physical or chemical weathering.

Anyway, biological weathering is very potent and soil formation would be impossible without it. Plants can be agents of mechanical weathering. Tree roots, which frequently are grow rock crevices, can exert sufficient pressure to force fragments apart (root pry). Some bacteria *(Bac. extorquens)* are capable to disintegrate and decompose silicates. Lichens, exuding organic acids, can dissolve feldspars. The residues of different forms of life are not less efficient in decomposing aluminosilicates and iron and aluminum oxides. Some burrowing animals like rodents, worms and ants bring to the surface rock particles to be exposed to the action of weathering.

All these organic activities taking place in the masses of disintegrated rocks signify the beginning of the primary process of soil formation. The roots and residues of higher plants accelerate such processes very much.

Primary minerals of igneous rocks (feldspar, amphiboles, pyroxenes, and micas) are transformed into secondary minerals (salts, oxides and hydroxides, clay minerals, etc.).

The upper part of the Earth’s crust, where the processes of weathering are noticeably active, is sometimes called the **cortex of weathering.** Soil scientists prefer to classify the cortex of weathering by dividing it into (1) sialitic; and (2) alitic.

Sialitic cortex of weathering is widespread in the regions with moderately humid climate and not so high temperatures. Clay minerals, resulting from weathering, mainly smectites and hydromicas, are resistant to further changes and tend to accumulate in this cortex:

$$2K[AlSi_3O_8]+4H_2O+2H^+ = \underset{Pyrophyllite}{Al_2(OH)_2[Si_4O_{10}]}+2K^+ +2H_2SiO_4$$

Primary minerals resistant to weathering also accumulate in parent materials and soils. Quartz becomes the most abundant mineral in sedimentary rocks, parent materials, and mineral soils. Ukraine is a region with dominant sialitic cortex of weathering.

Alitic cortex of weathering develops in humid and relatively hot climates of tropics and subtropics. Clay minerals and hydromicas cease to be resistant to weathering under such conditions and transform into hydroxides and oxides of Al and Fe.

Even kaolinite decomposes as a result of the formation of gibbsite and silica. Gibbsite is a mineral more resistant to weathering than kaolinite. The loss of silicic acid by leaching results in the progressive loss of silicon. There is a progressive increase in the accumulation of aluminum which tends to be incorporated into resistant secondary minerals that accumulate in the soil.

10.2. Soil-Forming Rocks (Parent Materials)

Soil is a disperse three-phase system. Some scientists consider it a four-phase system, adding to the three ordinary phases (solid, liquid, and gaseous) that of living organisms. The soil, therefore, is a heterogeneous, polyphasic, particulate, disperse, and porous system in which solid part occupies 40-60% of its volume and porosity may be not less. The solid phase constitutes the soil matrix. In mineral soils the solid part occupies 90-99% and more of their mass. The particles of the solid phase vary in chemical and mineralogical composition as well as in size, shape, and orientation, which, in their turn, depend on the composition of the initial soil-forming rock or parent material. Parent material is the unconsolidated mass in which a soil forms. It determines the limits of the chemical and mineralogical composition of soil as well as its texture or mechanical composition. There are five factors of soil formation identified by V.V.Dokuchaev: 1) parent material, 2) climate, 3) organisms (alive and dead), 4) landforms (relief), and 5) time. The Earth's crust consists of igneous, metamorphic, and sedimentary rocks.

Igneous rocks are those that form from the cooling and crystallization of hot silicate liquids. The silicate liquids themselves are the product of partial or complete melting of the crust or mantle. **Magma** is the term to describe the liquid within the Earth's crust or mantle. Lava is the term used to describe the liquid when it has reached the surface of the Earth. Sometimes magnetic materials are poured out upon the surface of the earth as when lava flows from a volcano. These molten materials are known as **eruptive, extrusive,** or **volcanic rocks**. Under other conditions, magma does not come to the surface but can force their way or intrude into other rocks where they solidity. These intruding rock materials harden and form **intrusive,** or **plutonic,** rocks, like granites, pegmatites, dunites, etc. Magmatic rocks take up 95% of the total weight of rocks in the Earth's crust but their soil-forming function is limited mainly to mountainous regions.

Metamorphic rocks form when already existing rocks are buried deep within the Earth and tremendous heat, great pressure, and chemical reactions may cause them to change into different rocks with different textures and structures (like slates and gneiss). Their importance for soil formation is very limited. The essential part of the Earth is covered by sedimentary rocks.

Sedimentary rocks are all rocks exposed on the surface of the Earth and subjected to processes of weathering, erosion, transportation and ultimately by deposition. Such processes produce sediment, which may eventually be transformed into sedimentary rock. The most widely used classification system for sedimentary rocks places them into three main categories according to origin of the materials from which they are made. These three categories are: clastic rocks, organic rocks, and chemical rocks. Sediments are typically deposited in layers or beds called strata. These rocks, represented by such common types as sandstone, shale, loess and limestone, make up about 75% of the rocks exposed on the Earth's surface. Nonconsolidated sedimentary rocks, made of the fragments of previously existing rocks, are the most important as parent materials of soils. Young sedimentary rocks were formed mainly in quaternary period as a result of weathering of solid rocks and translocation of the products by water, wind, and ice. Their formation continuous to this day. Being loose and porous they are favorable for soil formation. Their water and air holding capacity and penetrability as well as mineralogical, chemical, and mechanical composition determine the conditions of plant growth and soil-forming (pedogenic) processes.

10.3. Essential Genetic Types of Parent Materials

Parent materials are predominantly the sedimentary rocks of the quaternary period, only occasionally being of a more ancient origin. According to their genesis, the parent materials are divided into the following groups: eluvium, diluvium, colluvium, alluvium, moraine (glacial till), fluvioglacial (glacial outwash), lacustrine, marine, aeolian and other deposits.

Eluvium is represented by the products of rock weathering remaining at the place of their formation that is on the weathering bedrock covering it by eluvial coat. Eluvium is also called **the mantle rock**, or mantle-layers of loose weathered rock material, which lie in a relatively undisturbed position beneath the mantle, and deteriorate as it become exposed to the various agents of weathering. Eluvial rocks are mostly clastic sedimentary rocks with stones, rubble, and gravel, mostly angular fragments the size of which increases with depth. As parent material, eluvium occurs in mountain regions and on uplands where solid bedrocks appear on surface. Eluvium of granites, limestones, sandstones, and slates are common in the Crimean and Carpathian Mountains and on the Donetsky Upland. The soil formed in eluvium is skeletal and sometimes difficult or impossible to till. A parent material derived from sandstone is, essentially, composed of the sand particles that were cemented together in the sandstone. In the case of limestone, the soil often develops in the insoluble impurities that remain after the calcium and magnesium carbonates dissolve and leach from the weathering environment. Clay is common impurity in limestone, giving rise to the fine texture of soils derived from limestone weathering.

Most soils, however, have developed from sediments that were transported by water, wind, ice or gravity.

Deluvial deposits are the former eluvial ones transported and deposited along the slopes by surface runoff. Such sediments are layered and their particles are well sorted. Their thickness increases downslope. Sometimes the layering of their profiles is the inverted layering of the respective upland profiles.

Colluvial sediments occur at the base of steep slopes where gravity is the dominant force, causing movement and sedimentation. Colluvial sediments are common and are an important parent material in mountain areas. The particles of these clastic rocks are very unsorted and stony.

Talus slopes are a one form of colluvial deposits. Pieces of broken rocks pile up in these areas. These piles of rock fragments are called talus slopes.

Proluvial deposits form in mountain areas too. They form on the foothills as a result of temporary but intensive flow of water and mud. Sometimes gravity pulls loosened rocks down mountain cliffs in **a landslide**. A landslide is a large movement of loose rocks and soil. As the rocks fall, they collide with one another and break into smaller pieces. Falling rocks generally occur in areas where a road or highway has been cut through a rock formation, leaving cliffs on one or both sides of the road. Again the particles are very unsorted and stony.

Alluvium deposits are those left by flowing water. They occur in flood plains, terraces, fans, piedmonts, and deltas. Alluvial land is desirable for many purposes because it tends to have smooth topography, deep, fertile, permeable soils, and more readily available water than most other land.

Floodplains are narrow or nonexistent in young valleys but become broader and flatter as the landscape ages. Heavy runoff flow streams across these areas deposit a fresh layer of alluvium. The deposition produces a nearly flat surface except when stream channels cut the floodplain.

Alluvium in river valleys is usually only a few feet thick, although there are some exceptions. Terraces are formed when the base level of a stream changes and its channel is cut too deep for normal flooding to occur. River systems are complex systems and a catchment area will normally harbor a variety of landforms. These vary with the nature of the river and with the position in the river system. The stream may cut through a barrier to a lower base level, or uplift may increas the slope of the stream channels. Either cause leaves the former flood plain higher than the stream and its new floodplain. The old floodplain becomes a terrace.

Alluvial deposits contain materials which have been sorted according to size and as a consequence, are stratified with the coarser materials on the bottom. Alluvial materials may be composed of rock fragments, which have been smoothed or rounded by stream abrasion. The alluvial deposits are called **recent** if formed within the floodplains and **ancient** if formed on the terraces.

Lacustrine deposits, also called lake-laid clays, are composed mostly of fine sediment carried past deltas into the main body of water. The texture of the deposits may vary from silt or sand during seasons of high runoff to fine clay during cold or dry seasons, thus producing a layered effect. These layers are called verves and are sometimes counted to estimate the age of the deposit.

Glaciers crossing hilly topography produced many small lakes whose locations are marked by lacustrine deposits. They are also called glacial lacustrine deposits and are quite common in the Forest zone of Ukraine.

Products of Glacial Transportation and Deposition. Glaciers are large, slow-moving masses of land ice formed by the recrystallization of snow. The nature of glacial movement is not completely understood. Start their movement as the force of gravity and pressure from the weight of accumulating ice causes the ice in the lower levels of the glacier to become plastic and subject to slow flowage. Alternate periods of melting and refreezing produce contraction and expansion of the ice which may increase glacial movement. Glaciers are capable of carrying great quantities of earth materials and some of these rock fragments may be quite large. A glacier’s load will normally include finely pulverized rock flows and huge boulders, with all sizes of rocks in between. When the ice melts it will drop the debris, forming a variety of deposits, which are designated as glacial drift. There are two types of drift: till, which has not been stratified or sorted by water, and outwash, or stratified drift, composed of materials that have been sorted and deposited in certain layers by the action of glacial melt water. Deposits of till form topographic features known as moraines – ridges or mounds of boulders, gravel, sand, and clay deposited by a glacier. A terminal moraine, or end moraine, is a mound of till formed at the end of a glacier. Irregular deposits of till left by retreating melting glaciers or outwash left by streams of glacial melt water are known as glaciofluvial deposits. Outwash plains are broad, fan-shaped deposits of fine drift. Glacial outwash of sandy texture is the most widespread parent material in the Ukrainian Forest Zone.

Marine deposits. When waves or currents have reduced velocity they will deposit their load. Some rock particles will be thrown up on the shore by smashing. Most of the sediments thus deposited consist of rock fragments derived from the mechanical weathering of the continents, and they differ considerably from terrestrial or continental deposits. Beaches are coastal deposits of debris, which lie above the low-tide limit in the shore zone. Long narrow accumulations of sand lying parallel to the shore and separated from the shore by a shallow lagoon are called offshore or barrier bars, like Tendrivska and Obitochna bars of the Black Sea. Marine deposits may be saline and saline soils form in them.

Eolian (Wind) Deposits. The major types of wind-blown deposits are dunes and loess. Sand dunes are mounds or hills of sand which have been deposited by the wind. Dunes are formed in the areas where there is a sufficient amount of loose, unprotected sand and winds strong enough to move it. Areas of this type include sandy deserts, sandy floodplains, and sandy beaches along lakeshores or seacoasts. Small furrows, known as ripple marks, are commonly found on the windward slope of the dune. Crescent-shaped dunes in the deserts where winds blow steadily and from a single direction are called barchans.

Loess is the dominant parent material in Ukraine. Loess is wind-deposited material dominated by silt-size particles. Loess is well-sorted, usually calcareous, non-stratified, yellow-grey aeolian (eolian) clastic sediment. It consists predominantly of silt-sized particles, and contains normally less than 20% clay and less than 15% sand. Course silt makes up over 50% of its weight. There is no horizontal layering in loess, though there might be some in loess-like loams. The aeolian origin of loess is evidenced by the following facts:

1. Loess occurs as a blanket over a wide range of surfaces, to a large extent independent of topography;
2. Loess blankets are thicker on the leeward sides of obstacles;
3. There is absolutely no correlation between the mineralogy of the loess blanket and that of surface strata;
4. The grain-size distributions of loess are typical for the material transported in suspension.
5. Loess deposits are thicker near the source and become thinner with increasing distance;

6. Intercalated paleosols (fossil soils) are common in most loess belts;
7. Loess deposits is still found today around desert areas.

Chemical composition of Ukrainian loess is given in the following table.

Table 16. Chemical Composition of Ukrainian Loess (% by dry weight)

Components	Kiev Region	Kherson Region
SiO_2	73	66
Al_2O_3	6.9	15.2
Fe_2O_3	2.7	6.5
MgO	1.3	2.8
CaO	6.5	4.0
K_2O	1.6	2.1
Na_2O	1.0	1.1

Loess deposits and the soils formed in them are susceptible to both water erosion and mass movement. Falling raindrops and falling water can readily detach silt particles that are easily transported in runoff water. Gully erosion is relatively common in loess materials. The nearly vertical sides and flat bottoms of such gullies are described as U-shaped. Most other gullies are V-shaped. When an exposed loess surface becomes saturated with water, it is subject to mass movement ranging from small slumps to large landslides. But in other respects loess is an ideal parent material with good physical properties and favorable chemical composition.

Loess-like loams are result from loess transportation and deposition by water and are therefore layered in structure. The lower elements of topography are covered by them.

Organic deposits. In locations where considerable quantities of plant material grow and where decay is limited because of water saturation and oxygen deficit, low temperature, or acidity, an abundance of organic matter may accumulate. Sometimes this results in the formation of a mat of organic matter on the surface of mineral soil. Organic deposits are the parent material for organic soils. In permanently puddled areas thick deposits of peat and muck are formed.

10.4. Parent Materials of Ukrainian Soils

10.4.1. Parent Materials of the Forest Zone (Polissya)

The most common parent materials of the Forest zone are glacial till, glacial outwash and alluvial deposits. Less common are loess and loess-like loams, lacustrine, peat, chalky marl, and other deposits. Sand and loamy sand of fluvioglacial deposits (glacial outwash) are the predominant parent materials here, occupying 10.5% of the total area of Ukraine (60.3 mln hectares). The sand in them is coarse and medium grained, composed of quartz with rare admixtures of feldspars or hornblende. Sometimes small pebbles and gravel may be encountered in their layers. Fluvioglacial loams are very rare. The color of glacial outwash is all shades of yellow to yellowish grey. Very often they are underlain by moraine, igneous and solid sedimentary rocks like limestones and chalk. The most common in Polissya sod-podzolic soils (Albeluvisols, Spodosols) were formed in glacial outwash. Glacial outwash covers the plains with predominantly level topography. On elevated elements of topography (mounds, grades, and small hills) the parent material is represented by glacial till. Glacial till deposits make small "islands" on the uplands surrounded be the "sea" of glacial outwash. Moraines are of variegated texture, mineralogical and chemical composition, mostly of brown, reddish brown and yellow color. More productive soils form in carbonatic ($CaCO_3$ – containing) moraines.

Alluvial deposits occupy about 9% of the total area of Ukraine. They are intrazonal, unlike moraine and glacial outwash.

Recent alluvium was deposited by floodwater along the present-day streams in recent time. It is layered, predominantly sandy in texture with sand grains well washed off the Fe_2O_3 coatings. Ancient alluvium on the terraces of river valleys is predominantly coarse sand, so that the soils formed in it are good for pine forests and not good for agricultural crops because of their low water-holding capacity and poor fertility.

Eluvium of solid calcareous rocks takes up 2.3% of the total area of Ukraine. Limestones, chalk, marl and other sediments may be found among them. Eluvium of chalky marl containing 50 to 100% $CaCO_3$ can be found in the west and southwest of Polissya. Ukrainian rendzinas were formed in this parent material. They are stony and rich in $CaCO_3$ with neutral to slightly alkaline reactions, moderately good for wheat, rye, and sugar beat, but not good for apple orchards, potatoes, and flax.

Peats. Central parts of flood plains are sometimes covered with peat. As parent material, peat may be acidic, neutral, and alkaline. Acid one dominates in the Ukrainian Polissya (north-western and western parts). Those of the left-bank provinces, especially the so-called lowland province, have carbonatic and saline deposits of peat. Upland peat is a rarity. Lowland peats dominate all over the Polissya.

10.4.2. Parent Materials of the Forest-Steppe Zone

Loess and loess-like loams are the dominant parent materials of soils in the forest-Steppe zone. With the exception of floodplain sandy soils, light-textured soils on the terraces, severally eroded soils on the slopes, and the soils formed in the eluvium of solid igneous and sedimentary rocks, the bulk area of soils in the Forest-Steppe zone formed in loess and loess-like loamy deposits.

Loess in Ukraine is a parent material of soils occupying 74.8% of the country’s area. Sand and silt loams are the most abundant, but to the south and east the texture of the loess becomes finer and loamy clays and clay loams appear. Such parent materials are rich in $CaCO_3$, suitable for the accumulation of humus and formation of grainy “agronomically valuable” soil structure. Being an ideal parent material, loess, though, and soils formed in it, is susceptible to erosion caused by both water and wind. The coarse silt fraction (0.01-0.05 mm in size), which amounts to over 50% of the weight of loess, is to a great extent responsible for the development of gullies. On the right bank of the Dnipro gully erosion sometimes makes quite an ominous spectacle to see. Erosion damage to the ditches along the roadways can also be found.

A “single-loess” terrace on the left bank of the Dnipro is covered by a saline variety of loess-like loam. Sodium bicarbonate and even carbonate (which drastically increases soil alkalinity) are present among the soluble salts in local soils.

Typical chernozems (Mollisols, Chernozems Chernic) dominate over the entire Forest-Steppe zone. They occupy the leveled plateaus with a slow runoff and elevated loess terraces. Grey forest soils (Alfisols, Greyzems Haplic, Albeluvisols Haplic) dominate on “uplifted” well-drained watershed plateaus. Consolidated clays, alluvium, aeolian and some other deposits are much less common parent materials.

10.4.3. Parent Materials of the Steppe Zone

The Steppe zone of Ukraine is subdivided into two zones: Chernozemic Steppe and Arid Steppe. The Chernozemic Steppe zone is, in its turn, subdivided into two subzones: the Northern and the Southern.

The entire Steppe zone occupies an area of 25 million hectares (40% of the total area of Ukraine). Chernozemic steppe exceeds 20 millions hectares and Arid Steppe reaches 4.7 million hectares. Loess is the dominant parent material in the entire Steppe zone. Loesses and loess-like deposits are heavy, clay-like loams. To the south, their texture becomes finer. In the same direction, they increase their $CaCO_3$ content and salinity, and decrease the depth of gypsum accumulation. By landforms the Chernozemic Steppe zone is mostly level, though in the northern subzone, there are some uplifted areas 150-300 meters above sea level, including Donetsky Kryazh. There the parent materials are represented by eluvium and deluvium of crystalline igneous and consolidated sedimentary rocks, such as sandstones, limestones, slates, shalves, chalk, marl, etc. The soils formed in such materials are shallow, skeletal and subject to various types of soil erosion.

Loess and loess-like loams and clays of the Prychornomorska Lowland are saline with sulfates and chlorides. Red-brown clays underlying them and the limestones of the tertiary period underlying the clays may serve as parent materials on the slopes. But the most frequently occurring parent materials on the slopes of the ravines and dormant gullies are deluvium and colluvium of loess. In the Crimean Steppe, the Southern Chernozems form mosaic patterns with the Chernozems formed in the eluvium of consolidated lime-containing rocks and clays. Gleyed loess makes a parent material for the solod soils of the flat-bottomed depressions filled with water during certain periods of the year.

10.4.4. Parent Materials of the Carpathian and Crimean Mountain Regions

The Carpathian mountain region is situated in the westernmost part of Ukraine and includes the territories of the following oblasts: Zakarpatska (the entire territory), Ivano-Frankivska (the greater part), Lvivska (south), Chernivetska (west).

On the mountain slopes of the Carpathians, where acid brown soils formed, the essential parent materials are the eluvium and deluvium of sandstones and slates. These materials are noncalcareous, mainly sands, sandy and silt loams, skeletal to a varying extent. They contain rounded and angular fragments of solid rocks. Their permeability to water is very high. In the lower parts of the slopes and in the valleys of the mountain streams pebbles and conglomerates can be enconured. Ancient alluvial deposits, primarily loamy in texture, are found in the Transcarpathian regions on both sides of the Carpathian Mountain range. Alluvial deposits are also rich in pebbles and gravel.

Sheltered by the mountains from the north winds, the southern shore of the Crimea has a mild climate of the Mediterranean type. The cortex of weathering here is alitic (red-colored). The products of weathering of andesites and basalts, clayey shales, limestones and other solid rocks are rich in iron and aluminum oxides and hydroxides. Brown and red-brown xerophyte forest soils were formed in these parent materials.

On the mountain slopes, the parent materials are mainly calcareous. On the foothills of the Crimean mountains, the soils were formed in the eluvium of limestones, chalk, and marl. Such parent materials form skeletal soils. The soils on the slopes were formed in the eluvium of sandstones and shales (slates). These are brown forest soils with neutral soil reaction which makes them different from their Carpathian counterparts.

Questions and Assignments

1. Speak on physical (mechanical) weathering.
2. Cortex of weathering and its classification.
3. Name the essential types of chemical weathering, illustrating them with the equations of reactions.
4. Which rocks can be called parent materials?
5. Describe glacial till and glacial outwash.
6. Which parent materials are most common in the Forest zone of Ukraine?
7. What is loess and which properties does it impart to the soils formed in it?
8. Which parent materials are most common in the Steppe zone of Ukraine?
9. What is the difference between the loess and loess-like loam?
10. Describe alluvium as a parent material.

Chapter 11

Mineralogical and Mechanical Composition of Soils and Their Parent Materials

11.1. Soil Mineralogy

Soils and parent materials from which they were developed contain primary and secondary minerals. Primary minerals compose igneous rocks, whereas in unconsolidated secondary rocks and soils they are a residual material that avoided transformation caused by weathering and soil formation. Secondary minerals were formed from the primary ones under the influence of climatic and biological agents.

Primary minerals are represented mainly by the textural fraction over 0.001 mm in diameter, whereas secondary minerals – by clay particles less than 0.001 mm in size. In most soils, primary minerals dominate by weight over the secondary ones except in some feralitic soils. Primary minerals of soils are represented by quartz, feldspars, amphiboles, pyroxenes, and micas. These minerals compose the great bulk of igneous rocks, as is evident from the following Table17.

Table 17. Average Mineralogical Composition of Igneous and Sedimentary Rocks (data from Clarke, 1924)

Mineral Constituent	Origin	Igneous Rock, %	Slate, %	Sandstone, %
Feldspars	Primary	59.5	30.0	11.5
Amphiboles and Pyroxenes	Primary	16.8	–	small amount
Quartz	Primary	12.0	22.3	66.8
Micas	Primary	3.8	–	small amount
Titanium minerals	Primary	1.5	–	small amount
Apatite	Primary or secondary	0.6	–	small amount
Clays	Secondary	–	25.0	6.5
Iron oxides	Secondary	–	5.6	1.8
Carbonates	Secondary	–	5.7	11.1
Other minerals	–	5.8	11.4	2.2

As primary minerals have different resistance to weathering, their relative content in parent materials and soils is different from that in igneous rocks. The amount of quartz as most weathering-resistant mineral reaches 40-60%, and may be much higher. The second place is taken by the feldspars (up to 20%), represented mainly by the orthoclase $(KAlSi_3O_8)$, whereas $Ca-Na$ and $Na-Ca$ plagioclases are found much less frequently. Quartz and feldspars are found mainly as coarse grains, as their weathering proceeds slowly. They are concentrated mainly in sand and silt textural fractions of the soil or in parent materials. Amphiboles, pyroxenes and biotite micas are very susceptible to weathering and, therefore, are found in sedimentary rocks and soils in small amounts and in the form of tiny crystals. Particle size, through its effect on specific surface, is an important factor affecting the weathering rates of minerals. This mineral in the silt fraction weathers faster than if it is in the sand fraction. Chemical bonding within the mineral crystal is the major factor affecting weathering rate.

Oxygen, silicon, and aluminum are the three most abundant elements in the Earth's crust (47%, 28%, and 8% by weight, respectively). The silicon ion fits into an interstice formed by four oxygen atoms. The covalent bonding between O and Si forms a tetrahedron. Each tetrahedron $(SiO_4{}^{4-})$ has a net charge of -4. Uniting with each other through the oxygen ions, the tetrahedra arrange into individual, single-chain, double-chain, sheet, and 3-dimensional structures (Fig. 71).

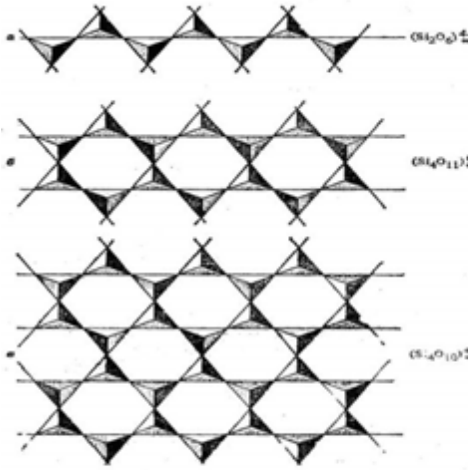


Fig 71. Common Arrangements of Silicon-Oxygen Tetrahedra in Silicate Minerals: a) single chain; b) double chain; c) sheet.

Individual or island type of structure is characteristic of olivine $[(Fe,Mg)_2SiO_4]$. Single-chain structure is characteristic of pyroxenes (augite) and double-chain one–for amphiboles (hornblende). Sheet structure is encountered not only in primary minerals (micas) but also in hydromicas and silicate clay minerals (kaolinite, smectite, etc) very common in heavy-textured soils. Each tetrahedron may be united with four others, forming framework or three-dimensional structure characteristic of quartz and feldspars, the minerals most resistant to weathering and, therefore, the most widespread in the earth's crust. In framework structures the silicon $[Si^{4+}]$ ion may be substituted by Al^{3+} and one free charge appearing at that is compensated by the cations of K^+, Na^+, and Ca^{2+}.

Aluminum (and magnesium or iron) octahedron is another structural unit of the crystal structure of silicates $[Al(OH)_6{}^{3-}]$. Primary minerals in the soils, especially those represented by a coarse-grain fraction determine agrophysical properties of soils and are a reserve source of the "ashy" elements of plant nutrition as well as a source for the formation of secondary minerals.

Secondary minerals of soils are usually considered in three groups: 1) ordinary salt minerals, 2) secondary oxides and hydroxides, and 3) silicate clay minerals.

Ordinary salt minerals are formed as a result of weathering and soil formation. They include calcite, magnesite, dolomite $[Ca,Mg]CO_3$, soda (Na_2CO_3), gypsum $(CaSO_4 \cdot 2H_2O)$, mirabilite, halite $(NaCl)$, phosphates, nitrates, etc. Such minerals are able to accumulate in large amounts in arid climates. They play a significant role in the development of soil salinity, its character and extent. Quartz is most mineral of oxides and hydroxides in soil environments. Disordered cristobalite commonly occurs in soils. Opal is a hydrated "amorphous" silica that comprises the bulk of diatomaceous rocks of biogenic origin. Quartz is an important part not only of the igneous but of many sedimentary rocks, parent materials and soils. Most sands are composed largely of quartz fragments. It often constitutes the major components not only of sand but also of silt and coarse clay fractions. Of iron oxides and hydroxides, hematite (Fe_2O_3) and limonite $(2Fe_2O_3 \cdot 3H_2O)$ may be very common in soils, but goethite $(FeOOH)$ is generally considered the most abundant. It is a significant component of most oxisols (red earths, krasnozems). Goethite and other Fe oxides containing Fe^{3+} are unstable in reducing environments. The solubilized Fe^{2+} may be translocated to another part of the profile, where there is a more oxidizing environment to form mottles or concretions. Limonite, or rather ferrihydrite, is associated with bog iron deposits and occurs in spodic (illuvial) horizons of sand-textured soils, like those of the Ukrainian Polissya.

Bauxite $(Al_2O_3 \cdot 2H_2O)$ is actually a mixture of minerals. It occurs in earthy, claylike masses, or in pisolitic form as rounded concretions in a claylike matrix. Bauxite origin is also connected with alitic cortex of weathering in humid tropics and subtropics Gibbsite, $Al(OH)_3$, is by far the most common Al hydroxide in soils. Sometimes it is the predominant mineral in the clay fraction. The usual path-way proposed for the gibbsite formation is by the desilication of kaolinite:

$$Al_4(OH)_8[Si_4O_{10}] + 10H_2O = 4Al(OH)_3 + 4Si(OH)_4$$

Boehmite $(AlOOH)$ is much rarer in soils than gibbsite. It is easier identified in intensively weathered soils. Silicate clay minerals are represented by montmorillonite group, kaolinite group, hydromicas, chlorites, and mixed-layer minerals. Their common feature is a sheet silicate structure. A typical layer silicate may be a combination of two types of structural units. One is a two-dimensional array of $Si-O$ tetraedra (the tetrahedral sheet). The second is a two-dimensional array of cations (generally Al^{3+}, Mg^{2+}, Fe^{2+}, or Fe^{3+}) in octahedral combination with oxygen or hydroxyl anions (the octahedral sheet). The tetrahedral and octahedral sheets are bonded together by a layer of shared oxygens. The size of layer silicate crystals is a direct consequence of the strength of their interlayer bonds. The degree of swelling is inversely related to the interlayer bonding energy. The surface area of layer silicates is related to their expanding properties, and may be either only external or both external and internal. External surface refers to the faces and edges of the whole crystal; internal surface refers to the area of basal plane surfaces. Nonexpanding minerals exhibit only external surface, but expanding minerals have both internal and external surface. The internal surface area is normally much larger than the external surfaces.

Adsorbed cations are held by layer silicates to balance the negative charge of the unit layer resulting from isomorphic substitution and from broken bonds on crystal edges. The magnitude of the exchange capacity of the crystal edge is related to the number of broken bonds, and therefore is a direct function of the crystal size.

Smectites (Montmorillonite) are 2:1 layer silicates. Because of low layer charge, smectites freely expand. The c-spacing varies with the exchangeable cation and the degree of interlayer solvation. Complete drying yields spacing from 0.95 to 1.0 mm, and full hydration can swell from the layer to a distance up to tens of nanometers. A typical half unit-cell formula for montmorilonite is $Na_x((Al_{2x}Mg_x)Si_4O_{10}(OH)_2)$ in which Na^+ is the charge–compensating exchangeable cation. Typical cation exchange capacities (CEC) for montmorilonite range from 800 to 1200 mmoles (+) kg^{-1} (80-120 meq/100 g). The CEC is only slightly pH-dependent. The low layer charge allows the mineral to expand freely, exposing both internal and external surfaces. Such expansion yields a total surface area from 600 to 800 $m^2 \times 10^3$ kg^{-1} (600-800 m^2/g), with as much as 80% of the total due to internal surfaces. Montmorillonite has high colloidal activity, including high plasticity and cohesion, and high swelling and shrinkage. Smectites are common in Vertisols. Soils of the southern chernozemic and arid steppe of Ukraine are particularly rich in montmorillonite.

Kaolinite $(Al_2Si_2O_5(OH)_4)$ is the layer silicate mineral that typifies the kaolins. Kaolinite occurs as a rule in soils as hexagonal crystals with an effective diameter of 0.2 to 2 µm. Reference to the formation of kaolinite was made in regard to the weathering of feldspars (kaolinization). Another mechanism for kaolinite formation is the partial disintegration of 2:1 clays (like montmorillonite). The stripping away of one $Si-O$ sheet, from 2:1 layer, converts the 2:1 layer into a 1:1 layer, typical of kaolinite. Most kaolin structural units are held together in the basal plane by hydrogen-bonding between oxygen ions of the tetrahedral sheet and hydroxyl ions of the octahedral sheet. This presents expansion (swelling) of the mineral beyond its basal spacing of 0.72 µm. Surface area is limited to external surfaces and hence is relatively small, ranging from 10 to 20 m^2g^{-1}. Kaolinite is coarse clay with low colloidal activity, including low plasticity and cohesion, and low swelling and shrinkage. Most of 1-10 mmoles (+) / 100 g CEC of kaoinite has been attributed to dissociation of OH^- groups on clay edges. The CEC of kaolinite is highly pH-dependent. The large size of kaolinite particles often gives soils high in kaolinite clay content good physical properties. Many of the kaolinite particles occur in the fraction of 0.0002 to 0.002 mm in diameter.

Hydromicas and products of their weathering – vermiculites are the most essential natural source of K^+ in soil solution.

Vermiculites occur extensively in soils. The layer structure of vermiculite resembles that of the mica from which the mineral is derived. Thus both trioctahedral and dioctahedral vermiculites exist. Weathering or alteration of the precursor micas has replaced the interlayer K^+ by (predominantly) exchangeable Mg^{2+} and has expanded the c-spacing to 1.4-1.5 nm in most cases.

The layer charge in vermiculites gives rise to a CEC from 1200 to 1500 mmoles (+) kg^{-1} (120-150 meq/100 g), which is considerably higher than the CEC of montmorillonite. CEC is only slightly pH-dependent. Vermiculite swells less than montmorillonite because of its higher layer charge. The mineral is nonswelling when saturated with ions. Such ions are commonly termed fixed and cannot be exchanged with ordinary salt solutions. Total surface areas of vermiculite, when not K^+ or NH_4^+ saturated, range from 600 to 800 $m^2 \cdot g^{-1}$.

Chlorites occur extensively in soils and are examples of 2:1:1 layer silicates. The positively charged "brucite" sheet sandwiched between negatively charged micalike sheets restricts swelling, decreases the effective surface area, and reduces the effective CEC of the mineral (10-40 meq/100 g). Total surface area ranges from 70 to 150 $m^2 \cdot g^{-1}$. Chlorite occurs as a rule in sedimentary rocks and in productive soils derived therefrom. Serpentine–derived soils contain chlorite and often are infertile because of their high magnesium and low calcium contents.

As soils weather, soil colloids become enriched in oxides, hydroxides, and hydroxides of Al Fe, and Ti. These materials range in degree of organization from amorphous to crystalline. Allophanes are among most important nonlayer silicate minerals of most salts.

Allophane is a general name for amorphous aluminosilicate gel. The composition of allophone varies widely but includes mostly hydrated Al_2O_3, Fe_2O_3, and SiO_2. Only minor amounts of Mg^{2+}, Ca^{2+}, K^+, and Na^+ are generally present. Allophane may have a high CEC in neutral to mildly alkaline solutions (of the order of 150 meq/100 g), but the measured CEC is highly dependant on pH. Allophane may have a high surface area (70 to 300 $m^2 \cdot g^{-1}$), but this property also varies widely with degree of crystallinity and pH.

Zeolites are the three-dimensional framework structures, like the feldspars, in which tetraedra are linked by sharing their vertices. Tetrahedra may be linked into 4-, 6-, 8-, or 12- membered rings, joined together less compactly than in the feldspars. Cavities within the structure contain loosely-held water molecules and charge-balancing cations that are freely exchangeable. Some zeolites, however, have cavities of smaller size that effectively prevent movement of large molecules, leading to the use of zeolites as "molecular sieves".

11.2. Mechanical Composition (Texture) of Soils and Their Parent Materials

Physically, a mineral soil is a porous mixture of inorganic particles, decaying organic matter, air, and water. There are two very important physical properties of soils: soil texture and soil structure. The term *soil texture* refers to the size range of particles in the soil and to the relative proportion of the various size-groups in the soil. Granulometric composition is a term most widely used in our country.

Soil structure is the arrangement of soil particles into groups or aggregates. The term soil texture carries both qualitative and quantitative connotations. Qualitatively, it represents the "feel" of the soil material, whether it is coarse and gritty or fine and smooth. Notes that "an experienced soil classifier can tell, by kneading or rubbing the moistened soil with his fingers, whether it is coarse-textured or fine textured and can also assess in a semiquantitative way to which of the several textural "classes" the particular soil might belong (D. Hillel, 1982). In a more rigorously quantitative sense, the term *soil texture* denotes the measured distribution of particles sizes or the proportions of the various size ranges of particles which occur in this kind of soil. By their origin the particles the accepted term in Ukraine is *mechanical elements* may be mineral, organic, and mineral-organic. They are represented by the fragments of rocks and mineral grains (primary and secondary), humus substances and the products of interaction between the mineral and organic substances.

The particles may exist free or aggregated. Large aggregates can disintegrated into smaller ones and into elementary mechanical particles with various agents. In microaggregates (< 0.25 mm in size) the particles are held together so tightly that a certain chemical treatment is needed to disintegrate them. H.D. Foth in his well-known book defines **soil texture as the relative proportions of sand, silt, and clay in a soil.**

Soil particles (or mechanical elements) are classified into soil separates by their size or effective diameter. According to N.A. Kachinsky, the following separates called *fractions* in Ukraine are identified:

1. Stones	> 3 mm	5. Clay: coarse	0.001-0.0005
2. Gravel	3-1	fine	0.0005-0.0001
3. Sand: coarse	1-0.5	6. Colloids:	< 0.0001
medium	0.5-0.25	7. Physical clay:	< 0.01
fine	0.25-0.05	8. Physical sand:	1-0.01 mm
4. Silt: coarse	0.05-0.01		
medium	0.01-0.005		
fine	0.005-0.001		

All soil particles exceeding 1 mm in size are called the skeletal part of the soil, whereas those smaller than 1 mm in size are called fine earth. Various separates effect the properties of soils and parent materials in a different way. This follows from different mineralogical and chemical composition of the separates as well as from their different physical and physico-chemical properties (Tables 18-20).

Table 18. Mineralogical Composition of Textural Fractions Belonging to Glaciofluvial Loam (A.A. Rode)

Fraction size, mm	Content of Primary Minerals, %				
	Quartz	Feldspars	Micas	Hornblende	Other minerals
1-0.25	86	14	–	–	–
0.25-0.05	81	12	–	4	3
0.05-0.01	72	15	7	2	4
0.01-0.005	63	8	21	5	3
< 0.005	10	10	67	7	6

Table 19. Chemical Composition of Textural Fractions Belonging to Light Grey Forest Soil (Alfisol, Greyzem) (data of N.A. Kachinsky)

Fraction Size, mm	Content, % of ignited weight					
	SiO_2	Fe_2O_3	Al_2O_3	MgO	K_2O	P_2O_5
0.05-0.01	85.91	2.45	5.92	0.57	1.44	Traces
0.01-0.005	84.14	4.17	5.88	0.67	1.59	Traces
0.005-0.001	73.44	4.89	14.73	2.72	1.70	0.33
< 0.001	59.86	8.32	23.05	4.03	2.36	0.44
Intact Soil	81.67	4.90	7.18	1.22	1.85	0.16

Table 20. Water-Related and Physical Properties of Textural Fractions Belonging To Parent Materials (V.V. Okhotin and V.G. Tkachuk)

Fraction Size, mm	Water-Related Properties			Swelling, % by volume	Physico-Mechanical Properties		
	Maximum molecular adsorption moisture	Infiltration Rate, cm/sec	Height of Cappillary Rise, cm		Plasticity		Shrinkage, % by volume
					Upper Index	Lower Index	
					% of Wetness		
3-2	0.2	0.5	0	–	Nonplastic		–
2-1.5	0.7	0.2	1.5-3	–	>>	>>	–
1.5-1.0	0.8	0.12	4.5	–	>>	>>	–
1.0-0.5	0.9	0.072	8.7	–	>>	>>	–
0.5-0.25	1.0	0.056	20-27	0.0	>>	>>	-
0.25-0.10	1.1	0.030	50	5	>>	>>	–
0.10-0.05	2.2	0.005	91	6	>>	>>	–
0.05-0.01	3.1	0.0004	200	16	>>	>>	–
0.01-0.005	15.9	–	–	105	40	28	–
0.005-0.001	31.0	–	–	160	48	30	4.0
< 0.001	–	–	-	405	87	34	8.2

Let us consider the character of each separate (fraction) in more detail. Stones (> 3 mm) are represented mainly by rock fragments.

Stonyness is a negative property of soils, as it hinders or makes impossible soil tillage, and is in the way of seedling emergence and plant growth.

Soil stonyness is evaluated and classified in dependence of stone (>3 mm) content (Table 21).

Table 21. Soil Classification by Stonyness (N.A.Kachinsky)

Content of Particles > 3 mm in ø, %	Extent of Stonyness	Type of Stonyness
< 0.5	Nonstony	Is identified by the Character of soil skeleton The soils may be bouldery, Pebbly, rubbly, etc.
0.5-5	Slightly Stony	
5-10	Moderately stony	
> 10	Very stony	

Gravel (3-1 mm) contains rock fragments and mineral grains. It is inert in relation to soil fertility but improves soil drainage (infiltration rate). With an excess of gravel particles the soils' water holding capacity is rather low for field crops. Gravel particles, as well as those of sand, hasten the wear of tillage implements. The admixture of gravel in some soils may be favored by some crops: vineyards are productive on skeletal loams with pebbles and gravel in them. If the field capacity of gravel-containing soil is below 3% by weight the soil is unfavorable for the cultivation of field crops.

Sand (1-0.05 mm) is composed of the fragments of primary minerals mainly those of quartz and feldspars. The fraction is endowed with high infiltration capacity but, unlike gravel, has certain water-holding capacity and measurable height of capillary rise. Some natural sands, particularly fine-grained ones, are usable for the cultivation of field crops, provided their field capacity is not less than 10 % by weight. Sands with water-holding capacity not less than 3 to 5 % are tolerable for some trees.

Coarse and medium silt (0.05-0.005 mm). The separate of coarse silt with effective diameter within 0.05-0.01 mm is characterized by an increased content of micas which impart to it an increased plasticity and noticeable resistance to tillage. Medium silt has a higher water-holding capacity but its infiltrability is low. It has no capacity for flocculation and does not "participate" in the processes of soil structure formation as well as physico-chemical processes occurring in the soil. The soils rich in coarse and medium silt fractions are easily pulverized, prone to puddling and compaction, and their infiltration capacity is comparatively low.

Fine silt (0.005-0.001 mm) is composed of primary and secondary minerals. It has a capacity for flocculation (coagulation) and is useful for the formation of soil structure. It has adsorbing and cation (and anion) exchange capacity and contains an increased amount of soil humus. But the abundance of fine silt in the soil in a free (nonaggregated) state imparts to the soil a number of unfavorable properties including low infiltration rate, high content of unavailable water, well–marked swelling and shrinkage, cohesiveness, the propensity to form cracks at drying, and compacted constitution with low porosity.

Clay (< 0.001 mm) is composed mainly of secondary minerals. Primary minerals are usually represented by quartz, orthoclase, and muscovite. Clay fraction is very important for the development of soil fertility.

It plays the most significant role in the physical and chemical processes occurring in the soil. Its adsorbing capacity is very high, including CEC. It contains the maximum amount of humus compared with less fine fraction of soil textural composition, as well as available forms of plant nutrients. Colloidal separate within the fraction of clay is the most potent substance for the formation of good for agriculture preferable soil structure.

Water-related physical and physico-mechanical properties of soils rich in clay are to a considerable extent determined by its ability to coagulate and glue the mechanical particles of the soil into the aggregates of soil structure. This ability is, in its turn, determined by the mineralogical and chemical composition of the soil, great amount of organic matter in it (especially colloidal humus), and some amount the content of calcium and iron among the exchangeable cations. A well structured soil is characterized by favorable physical properties even with a high content of clay. Deflocculated or dispersed clay fraction has negative physical properties.

The properties of textural separates drastically change during he transition from the fraction of physical sand (> 0.01 mm) to that of physical clay (< 0.01 mm). Classifications of soil separates by the United States Department of Agriculture and International Soil Science Society Systems are given below.

Table 22. Classifications of Soil Separates and Their Surface Area

Separate	Diameter, mm		Surface Area in 1 Gram, cm^2
	USDA Classification System	ISSS Classification System	
Very Coarse Sand	2.00-1.00	–	11
Coarse sand	1.00-0.50	2.00-0.20	23
Medium sand	0.50-0.25	–	45
Fine sand	0.25-0.10	0.20–0.02	91
Very fine sand	0.10-0.05	–	270
Silt	0.05-0.002	0.02-0.002	454
Clay	Below 0.002	Below 0.002	8 000 000

N.A. Kachinsky's Textural Classification of Soils and Sedimentary Rocks is still standard and most popular in Ukraine. Soil division into textural classes is carried out considering on the ratio between the contents of two fractions: physical clay and physical sand. Count is also taken of the predominant type of soil formation, whether podzolic, chernozemic (Steppe), or sodic (sodic or severely sodicity-affected soils). The textural classes (or groups, according to N.A. Kachinsky's terminology) are given in the following table 23.

N.A. Kachinsky's textural classification was developed on the basis of soil genesis. Clayey properties of different soils with equal content of physical clay may be evident and display themselves to different extent. Steppe (chernozems) and humid subtropics (redzems and yellowzems) soils are more structured and make a transition to the category of heavier soils with higher content of physical clay than sodic and podzolic soils, which are less structured and possess some unfavorable "clayey" properties with a lower content of physical clay.

Table 23. Kachinsky's Textural Classification of Soils and Rocks

Soil Textural Class (Group) Name	Content of physical clay (<0.01 mm), %			Content of physical sand (>0.01 mm), %		
	Soil formation type			Soil formation type		
	Podzolic	Steppe chernozems + redzems and yellowzems	Sodic (solonetzic)	Podzolic	Steppe chernozems + redzems and yellowzems	Sodic (solonetzic)
Sand: loose	0-5	0-5	0-5	100-95	100-95	100-95
bound	5-10	5-10	5-10	95-90	95-90	95-90
Loamy sand	10-20	10-20	10-15	90-80	90-80	90-85
Loam: light	20-30	20-30	15-20	80-70	80-70	85-80
medium	30-40	30-45	20-30	70-60	70-55	80-70
heavy	40-50	45-60	30-40	60-50	55-40	70-60
Clay: light	50-65	60-75	40-50	50-35	40-25	60-50
medium	65-80	75-85	50-65	30-20	25-15	50-35

In the USA a textural triangle is used to determine a soil textural class. The textural triangle shows the limits of sand, silt, and clay contents of various texture classes. When the percentages of sand, silt, and clay are determined, the soil can be attributed to one of 12 major textural classes (Fig. 24).

The texture of a soil is expressed with the use of class names. The sum of the percentage of sand, silt, and clay at any point in the triangle is 100. The soil containing equal amounts of sand, silt, and clay is a clay loam. The area outlined by the bold lines in the triangle defines a certain class. The class of soils called loam occupies a location close to the centre in the textural triangle. Such soil contains a "balanced" mixture of coarse and fine particles. As such, loam is often considered the optimal soil for plant growth and for agricultural production, as its capacity to retain water and nutrients is higher than that of sand while its drainage, aeration, and tillage properties are more favorable than those of clay. D. Hillel (1982) in *Introduction to Soil Physics* notes that such a view is an oversimplification, as under different environmental conditions and for different plant species sand or clay can be more suitable than a loam.

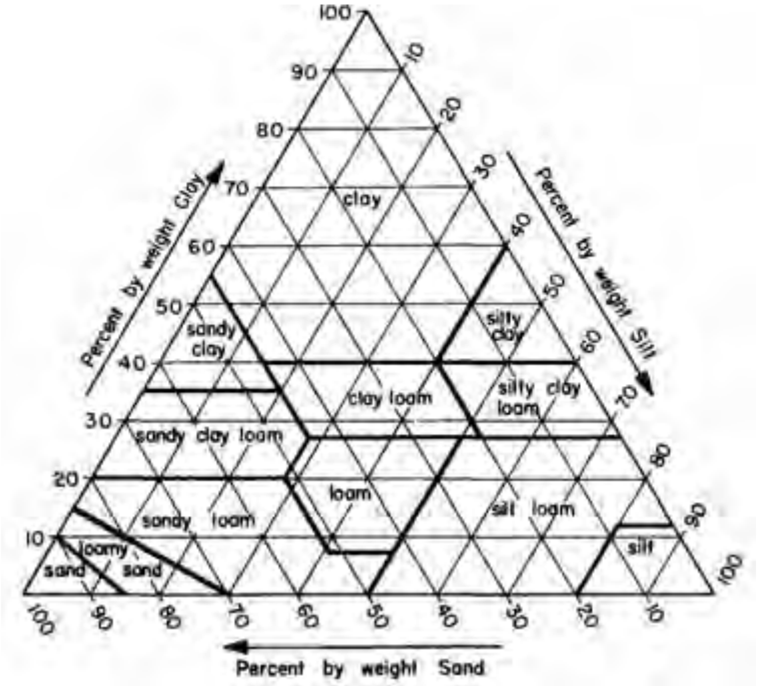


Fig 72. Textural triangle, showing the percentages of clay (below 0.002 mm), silt (0.002-0.05 mm), and sand (0.05-2.0 mm) in the basic soil textural classes.

Mechanical (Particle Size) Analysis

Sieves can be used to separate and determine the content of relatively large particles of the sand and silt separates. Sieves, however, are not recommended for the separation of the clay particles from the silt and sand. A sample of air-dry soil is mixed with a dispersing agent (such as sodium pyrophosphate solution). Then the soil-water suspension is stirred on a mixer for several minutes to bring about the separation of the sand, silt and clay particles. The suspension is poured into a graduated cylinder, and distilled water is added to bring the contents up to volume.

The soil particles settle in the solution at a speed directly related to the square of their diameter and inversely related to the viscosity of the liquid:

$$V = \frac{2}{9} r^2 \frac{\rho_s - \rho_e}{\eta} g, \text{ or} \qquad (13)$$

where r is the radius of a particle; η is the viscosity of the liquid; ρ_S is the density of a particle, ρ_1 is the density of the liquid, and g is the acceleration of gravity. A hand stirrer is used to suspend the soil particles thoroughly and the time is immediately noted:

$$t = 18h\eta / d^2 g(\rho_s - \rho_e) \qquad (14)$$

where h is the height of particle falling through the time t; d = diameter of particle.

One way of measuring particle-size distribution is to use a pipette (as it is used in Ukraine) to draw samples of known volume from a given depth (h) in the suspension at regular times after sedimentation began. An alternative method is to use a hydrometer (as it is used in the US) to measure the density of the suspension at a determined depth as a function of time.

The described procedure based on the use of Stokes' law is dependent upon certain simplifying assumptions which may not be in accord with reality. Among these are the following:

1) the particles are sufficiently large to be unaffected by the thermal (Brownian) motion of the fluid molecules;
2) the particles are rigid, spherical, and smooth;
3) all particles have the same density;
4) the suspension is sufficiently dilute that particles do not interfere with one another and each settles independently;
5) the flow of the fluid around the particles is laminar, i.e., no particle exceeds the critical velocity for the onset of turbulence;
6) soil particles, while indeed rigid, are neither spherical nor smooth, and some may have the form of a plate. Soil particles are not of the same density.

For all these reasons, the mechanical analysis of soils yields only approximate results.

Determining Texture by the Field Method. When investigating a land-use problem, the ability to estimate soil texture on location is useful in diagnosing the problem and in formulating a solution. A small quantity of soil is moistened with water and kneaded to the consistency of putty to determine how well the soil forms balls and wires. Sandy soils do not even form a ball (2-3 cm in diameter). Loamy sands form easily breakable balls but do not mold into wires. Wires of sandy loams are thick (> 4 mm) and break into pieces during the process of molding. Clay loams (medium loams, in N.A. Kachinsky's classification) may be shaped into wires which break into pieces in at attempt to make a ring round the finger. Loamy clays (heavy loams, in N.A. Kachinsky's classification) yield to making rings out of them with cracks on the outer surface. But they do not break into pieces. Clays “make rings” without cracks and it is possible to mold some clays into thin wires (2-3 mm).

Importance of Soil Texture. Soil texture greatly effects soil formation and the use of soils in agriculture and other branches of economy. Soil and parent material texture determines the intensity of soil forming processes connected with transformation, translocation, and accumulation of organic and mineral compounds in the soil. As a result, under the same environmental conditions, the soils with different properties are formed in parent materials of different texture.

Soil texture significantly effects soil water potential, water retention, height of capillary rise, character of water movement in saturated and unsaturated soil, availability of soil water to plants, physico-mechanical (related to soil tillage) properties of soils, soil air and aeration, heat properties and heat transfer in the soil, red-ox conditions of soil environment, soil adsorption and CEC, organic matter accumulation, as well as content and availability of nutrients to plants.

Soil texture determines the conditions and dates of soil tillage, fertilizers and soil amendments, and techniques of their application, Light-textured soils are easily put to tillage, have sufficient infiltrability, favorable aeration, and become warmer earlier in spring. But they possess some unfavorable properties and, first of all, low water holding capacity. They are poor in OM content and fertility (amount and availability of nutrients). Their CEC is low, and they are suffer from erosion by wind.

Heavy-textured soils possess unfavorable physical and physico-mechanical properties. Their infiltration rate is low. They easily puddle and form crust. They are called heavy because their resistance to tillage is large. But their water-holding capacity is high and they are better supplied with OM and the elements of plant nutrition.

Medium loams are endowed with the most favorable set of properties for most field crops. But even heavier soils may be good in many respects if they are well-structured and contain a sufficient amount of SOM.

Many clay soils expand and shrink with wetting and drying, causing cracks in walls and foundations of buildings.

Maximum productivity for wheat and other small grain crops in Ukraine is registered for medium textured soils (light and medium loams). Loam soils have the greatest productivity for corn and sugar beet. Sandy soils are good for tobacco. Vines like skeletal loams with pebbles and gravel. Potatoes prefer sands and loamy sands, and so do the strawberries.

Questions and Assignments

1. Name the differences between the mineralogical compositions of igneous and sedimentary rocks.
2. Name the primary minerals of soils and speak on their importance for soil productivity.
3. Classify the secondary minerals of soils and describe their formation.
4. Characterize silicate clay minerals. What properties do they impart to the soils?
5. Speak on vermiculites and their role in the formation of soil fertility.
6. Characterize allophone and zeolites.
7. Define the term “soil texture”.
8. Speak on the properties of soil separates.
9. Characterize N.A. Kachinsky’s textural classification.
10. Compare the classifications made by N.A. Kachinskyy and by USDA Classification systems for soil separates.
11. Describe the procedure of mechanical (particle size) analysis based on Stokes’ law.
12. Speak on the field methods of soil texture determination.
13. Give reasons for the importance of soil texture in crop production.
14. How does soil texture effect soil formation and properties?

Chapter 12

General Scheme of Soil Formation and Development of Soil Profile

The formation of a soil is the result of interaction of five soil-forming factors: climate, organisms, parent material, topographic position or slope, and time. Human beings and their activity is an additional, sixth factor of soil formation, which is becoming more and more dramatic as we now, hopefully, have passed the inflexion point on the world population growth curve which designates the end of the exponential stage of this growth. But the appetites and demands of human beings probably did not reach the inflexion point of their growth so far, and we, therefore, have to understand and predict the evolution of environmental changes including the changes in soils which still sustain life on earth.

12.1. General Scheme of Soil Formation

Soil genesis deals with factors and processes of soil formation. Soils are products of evolution, and soil properties are the function of time or soil age. The diversity of soil on the globe is tremendous. It is partly reflected in the complexities of soils classification and taxonomic units. But despite this tremendous diversity, there exists a general scheme of the soil formation process. The dominant components of the soil formation process are:

1) the transformation of minerals in a parent material and subsequently in the soil developed from it;
2) the accumulation of organic residues and their gradual transformation;
3) interactions between the mineral and organic compounds resulting in the formation of a complex system of organic-mineral substances;
4) the accumulation of biologically important elements, primary the elements of plant nutrition, in the upper portion of the profile;
5) the vertical movement of soil formation products within the profile, including leaching and the removal of soluble materials, the translocation of colloids, the capillary rise of ground water accompanied by the accumulation of soluble salts, etc.

Present-day American soil science identifies four groups of processes which form the soils as natural (or human-modified) bodies: (1) transformation, (2) translocation, (3) additions, and (4) losses.

The net effect of these processes is the development of soil horizons, that is, the genesis of a soil. All these processes are the result of interaction between great (geologic) and small (biologic) cycles of matter and energy on the globe. According to A.A. Rode, a noted Soviet pedologist, “soil formation amounts to continuous transfer and transformation of matter and energy within the profile of a soil” .

Soil formation, in addition to its “general scheme”, is characterized by evolutionary stages, types and regimes. It begins with biological weathering and develops through a of dialectically opposite processes and phenomena till it reaches the stage of equilibrium with the factors of soil formation, which, however, is not everlasting. Any disturbance of the equilibrium brings about a new evolution in soil formation.

12.2. Factors of Soil Formation

Role of Climate in Soil Genesis. Climate greatly affects the rate of soil genesis. For example, in areas, permanently hot and dry, as well as in those permanently frozen, soil does not form. The two components of climate very important for soil formation are precipitation and temperature. A continuous exchange of heat and moisture takes place between the soil and the atmosphere. This exchange forms a definite hydrothermal regime of soil. Thermal groups of climate were identified by the sum of mean diurnal temperatures exceeding 10ºC during the period of crop growth and development:

Climate Group:	∑ + >10ºC
Cold (Polar)	< 600ºC
Moderately Cold (Boreal)	600-2000ºC
Moderately Warm (Subboreal)	2000-3800ºC
Warm (Subtropical)	3800-8000ºC
Hot (Tropical)	> 8000ºC

The climates of these thermal groups are laid out like latitudinal belts surrounding the globe. The belts are characterized by certain types of vegetation and soils. They are referred to as “soil-bioclimatic belts”. Vysotsky-Ivanov humidity coefficient which is equal to the ratio between the annual precipitation and annual evaporation from the open water surface in a given location is used as a criterion to classify the climates into humidity groups:

Climate Group:	Humidity Coefficient:
Extra Humid	> 1.33
Humid	1.33-1.0
Semihumid	1.0-0.55
Semiarid	0.55-0.33
Arid	0.33-0.12
Extra Arid	< 0.12

Every 10ºC increase in temperature approximately doubles the rate of chemical reactions. Increased weathering and clay formation occur with an average increase in soil temperature. Water is also necessary for mineral weathering and plants growth conditions. The relationship between average temperatures and plants growth and the accumulation of organic matter is complex. The organic matter content of soil is the net result of plants growth or the addition of organic matter, the rate of organic matter decomposition, and the soil's capacity to protect organic matter from mineralization. Water in excess of field capacity (surplus water) participates in the downward translocation of colloidal particles and leaching of soluble salts. The limited supply of water in deserts results in soils that tend to be saline, alkaline, relatively unweathered, and low in clay, organic matter content, and CEC. Even slower are the rates of microbial decomposition of organic matter due to low temperature and soil wetness. The grassland soils tend to have a gradual decrease in organic matter with annual temperature increase.

Where there is sufficient water for only limited leaching, the carbonates tend to move downward only a certain distance, where they accumulate and form a horizon of calcium carbonate accumulation. Increases in precipitation are positively related with greater leaching of lime and greater depth to a carbonate layer, the development of soil acidity, weathering and clay content, plant growth and organic matter content.

Some soils have been influenced by more than one type of climate because of climate shifts. The processes of wind and water erosion are greatly affected by climate.

Organisms as a Factor of Soil Formation

The organisms make up the so-called biological factor of soil formation, which implies the versatile participation of living organisms and the products of their life cycles in soil formation. The characterization of various groups of soil organisms, participating in soil formation and the development of soil productivity will be considered in the next chapter, whereas here we shall briefly state that plants affect soil genesis by the production of organic matter, nutrient cycling, and the movement of water through the hydrologic cycle. Microorganisms play an important role in organic matter decomposition (and mineralization) and the formation of humus. Soil animals are consumers and decomposers of organic matter. However, some of them play also the role of earthmovers. To a first approximation, the most obvious effect of organisms on soil genesis is that caused by natural vegetation: whether it is trees or grass. The amount, quality, and distribution of organic matter is related to the differences in the growth habits of these two kinds of plants. Many students of soil formation believe that soil genesis is accelerated by woodland rather than grassland vegetation, but grassland soils have much greater content of organic matter within the soil than forest soils, and the organic matter diminishes less rapidly with the increasing soil depth.

Role of Parent Material in Soil Formation

Parent materials greatly influence the properties of young soils, including soil color, texture, structure, mineralogy, chemistry, and pH. Over time the effects of the parent material decrease, but soil texture, physical properties, mineralogical and chemical composition still remain dependent on the parent material.

Consolidated rocks are not parent materials but serve as their source. Soil formation begins not immediately after the deposition of volcanic ash or loess but after physical disintegration of hard rocks, when granite, basalt, or sandstone are exposed to weathering. During the early stages of rock weathering and soil formation, the formation of parent material and soil may occur simultaneously as two overlapping processes. Most soils developed from sediments transported by water, wind, ice, or gravity.

Under the same conditions of climate and topography, different soils can form from different parent materials. On the loess islands of the Ukrainian Polissya the soils are never soddy-podzolic like those formed from glacial outwash or alluvial sandy deposits. Grey forest soils typical of the Forest-Steppe zone are mainly formed within such “islands”. The salinity of parent materials in the southern zones causes the salinity and even sodicity of soils formed from such parent materials. Clay is a common impurity in limestone, giving rise to the fine texture of soils derived from limestone weathering. The soils formed from the weathering of cherty limestone are stony. Alluvial parent materials typically are stratified. Where streams dissect the landscape, various materials may be exposed for soil formation, giving birth to a number of soils whose differences are due mainly to parent material differences.

In permanently ponded areas throughout the world, thick deposits of peat and muck have been formed. Organic deposits are the parent material for organic soils.

Role of Topography in Soil Formation

Landforms may be of different origin – tectonic, glacial-accumulative, glacioerosional, eolic, alluvial, etc. Landforms are studied by the science of geomorphology. To understand the role of landforms in soil formation it is a tradition in our geomorphology to classify the relief of the surface into macro-, meso-, micro-, and nanoforms. **By macrorelief** the largest landforms are meant, such as plainlands, plateaus, and mountain systems. The origin of these forms is linked with tectonic phenomena within the earth's crust. **Mesorelief** is represented by the landforms of moderate size – hills, knolls, ravines, river valleys, terraces, etc. The origin of these forms is connected with the exogenous geologic processes (denudation, formation of continental sediments, etc.) which can be significantly effected by the slow movements of the earth's crust. The layers of the Earth's crust throughout its long history were lifted up, pushed down, folded, and faulted by tectonic forces. The resulting movements of the earth's surface are usually too small and too slow to be directly observed, but they are constantly changing the appearance of the land.

Microrelief is a term denoting some small land-forms from several square meters in area, with altitude ranges within one metre. Some small knolls, cavities, erosion rills, channels, etc are the forms of a microrelief.

Nanorelief is created by tillage operations and implements. The edges and beds of small streamlets (rill erosion) are also characterized as nanorelief. All depends upon the extent of soil disturbance.

Relief plays its soil-modifying role redistributing solar radiation and precipitation depending on the aspect and steepness of slopes and thereby effects water-related, heat, nutritive, redox, and salinity regimes of the soil.

Mountains cause **the vertical zonality** of climate, vegetation, and soils. Mountain ranges modify the climatic conditions in a locality.

Differences in topography can cause wide variations in soils within the confines of a single field. Topography determines the local distribution or disposal of the precipitation and determines the extent to which watertables influence soil genesis. The soils in this respect are classified in our country into three groups: 1) **automorphic,** formed on level surfaces and slopes with free surface runoff and depth of water table over 6 meters; 2) **semihydromorphic**, formed with short-time stagnation (ponding) of surface water or with the watertable within 3-6 meters deep (capillary fringe may reach the roots of plants); 3) **hydromorphic**, formed under conditions of prolonged ponding with surface water or with watertable no deeper than 3 meters from the surface (capillary fringe may reach soil surface). Water-permeable soils on broad, level areas receive and infiltrate almost all of precipitation. Soils on sloping areas infiltrate less than the normal precipitation; hence, there is runoff. Both the length and steepness of slope affects soil genesis. As the steepness of slope increases, there is greater water runoff and soil erosion. Generally, an increase in slope gradient is associated with less plant growth and organic matter content, less leaching and eluviation. Soils have thinner sola and are less developed on steeper slopes. Many soils on sloping land are in equilibrium in terms of erosion rate and the rate of horizon formation.

Erosion may not be the only reason for the soils on the slopes to have thinner sola. Surface runoff mainly occurs along the lines of preferential flows ("runoff papillars", according to M.I.Polupan, et al. (2005). Slope soils are more xeromorphic in comparison with those on the level areas as less water penetrates into them and they, therefore, are wetted to a lesser depth.

Role of Time in Soil Formation

Soils are products of evolution and soil properties are a function of time or soil age. When studying time as a soil-forming factor, the other soil-forming factors should be considered as constants or almost in variable values. Every new cycle of soil formation (seasonal, annual, perennial) brings about certain changes in the transformation of organic and mineral matter in the profile. The students of soil evolution usually make discrimination between the absolute and the relative ages of soil. Absolute age is a period of time expired since the beginning of soil formation to the present time. It may be only a few years or millions of years in length. Tropical soils are the most ancient as they suffered no dramatic disturbances (erosion by water and wind, etc.). The northern part of Ukraine saw the end of glaciations about 10 000 years ago. For a short period of time after the glacial ice melted, a tundra climate probably existed. The youngest soils can be found on the flood plain areas. As the soils evolved, soil properties changed and that caused a change in the composition of plant species and other organisms. Age sequences of soils may be studied on the terraces of some river valleys.

The rate of soil formation in any particular case is highly dependent on the nature of soil forming factors. Some soil features or properties develop quickly, but the development of other properties requires much more time.

Relative soil age characterizes the rate of soil formation and pertains to soil development sequences, or stages. H.D.Foth in his *Fundamentals of Soil Science* states that A horizon (surface horizon) can develop in a few decades. Consequently, soils with only A and C (parent material) horizons can develop in 100 years or less. The formation of a B horizon involves changes in color and/or structure and requires about 100 to 1000 years. The time to form intensively weathered soils containing a great amount of kaolinite and oxides of iron and aluminum is approximately 100 000 years. Some of these soils in the tropics are more than a million years old" [14].

Human Being as a Soil-Forming Factor

The use of land for agriculture, forestry, grazing, and urbanization produce extensive changes in soils. Soil tillage, fertilizing, reclamation, etc. affect not only the soil but the entire complex of environmental conditions. Humans in their activity try conscienciously and with a definite aim in mind to reconstruct soil body and change its properties and regimes to make the soil more productive. Drainage and irrigation can change the soils much more rapidly than the natural factors of soil formation. Sometimes the soils become better, particularly those with low natural fertility and plant productivity. Such soils are then called "cultured". Very often the soils and the entire natural environment become worse. Such phenomena are termed *soil and environmental degradation*. Sometimes, solid waste disposal produces new soils that are not the same as just the accumulation of trash. But millions of hectares of land have soil properties that are due more to human activities than to natural soil-forming factors. H.D.Foth (1990) [14]. notes the fact that "in China land has essentially been created to grow food crops". The enormous movement of soil and the shaping of the landscape were required in China for growing paddy rice. V.M.Starodubtsev, et al (2005), analyzing soil desertification in the river deltas, identified some law-governed and predictable "patterns" in soil evolution which should be taken into account in the development of land use (and natural resources use) projects for the environmental protection.

12.3. Soil Development and Evolution

Factors of soil formation have specific effects on soil genesis and cannot be substituted by each other. In this respect they are equisignificant. But their effects are displayed through their interactions. An example of topography, parent material, and time interactions may be the following: on slopes, water runoff and soil erosion occur while simultaneously run-on water and soil deposition occur on areas at a lower elevation. Young soils tend to occur where both sedimentation and erosion are active. The oldest soils tend to occur on broad upland areas where neither erosion nor sedimentation take place.

The genesis of any soil is composed of at least three consecutive stages: 1) inception of soil formation, or primary pedogenic process; 2) contradictory development of soil as a natural body; 3) formation of a mature soil approaching the state of equilibrium with the environmental factors (or, rather, the factors of soil formation).

The beginning (inception) of soil formation coincides with the beginning of the surface ecosystems functioning under conditions of continuous interaction of the five factors of soil formation, when ever the parent material reaches the day-light surface of the earth. The first stage of soil formation on the surface of solid rocks known as the *primary soil formation*, usually lasts for a very long time. Rocky materials need some disintegration and chemical transformation before they can sustain life cycles of even the “pioneer” species, like lichens and mosses. M.V.Lomonosov described it in the following words: “and even the barren rocky mountains show on them the greenery of young moss which subsequently darkens and turns into earth. The earth, having accumulated by the duration of time, serves than for the growth of some larger mosses and other plants” [14].

Alongside the processes occurring within the frames of biologic cycle, some abiotic processes of various nature (physical, physico-chemical, and chemical) also take place. These are the processes of dissolution, precipitation, evaporation, condensation, sorption, diffusion, complex formation, etc. Such processes may occur not only in soils but also in other environments. They are not specifically soil processes. Any of them taken up separately does not form any specific soil features, although they are actively participating in all soils and in all stages of soil formation. A.A.Rode proposed to identify them as *microprocesses.* At the initial stage of soil formation these processes are not yet united into a single system and have mostly an independent character of each other. But at the final substage of a primary stage the microprocesses eventually become closely interrelated. Abiotic and biotic processes become “cause and effect”. Some definite biogeochemical cycles become increasingly distinct in the young soil and specific soil features appear in the profile.

As the rate and magnitude of biologic cycle of matter and energy in the forming soil dramatically increase, the stage of soil development sets in. Any nutrient element returns into the soil after organic matter decomposition in the form of new compounds not formerly contained in the soil or parent material. A certain pool of labile substances (reserve pool) is formed in the soil which contains available nutrients in excess of soil biota needs which are equal to the so-called exchange pool. Soil physical properties also change and become more favorable for plant growth, improving soil porosity (aeration porosity), water retention and percolation as well as penetrability and “comfort” for the roots of higher plants, the main producers of soil ecological system. Solid phase becomes more or less aggregated. Vertical transfer and translocation of the products of soil formation brings about creation a specific profile. In this stage of soil development, the magnitudes of biologic and abiotic cycles reach a greater equilibrium with the factors of soil formation. A qualitatively new pedogenic processes develop in the soil profile, the so-called according to (A.A. Rode’s term *mesoprocesses*) of soil formation. Mesoprocesses form specific soil properties. They are very numerous and include podzolization, humus accumulation, development of sodicity, peat accumulation, lessivage, aggregate formation, etc. Soil body acquires a spatial differentiation of composition and properties at the levels of soil aggregate (microaggregate) and soil horizon, which determine the general character of soil profiles.

Soil macroprocesses (according to A.A. Rode), or the soil-forming processes per se, bring about the formation of specific soil types, not just specific soil properties and horizons. Each of soil types has a definite system of soil horizons like chernozems (mollisols), podzolic soils (spodosols), etc. Soil macroprocesses are a set of appropriate mesoprocesses occurring under specific conditions of biogeochemic cycles. Higher plants are most powerful translocators of biogenic and other absorbed elements from the various horizons of soil profile and parent materials to the surface horizon. The stage of soil development may last for hundreds, thousands and more years. Factors of soil formation (climate, for instance) may undergo changes during such long periods of time. Under such conditions, a situation may arise when the soil having achieved the equilibrium state by some soil features may "transit" to a new state of unequilibrium and reach a new stage of its development. Soil profile may therefore preserve the features of the preceding conditions of soil formation, which, however, are alien to the new conditions. An example may be the formation of "repeatedly" saturated with calcium carbonate "regraded" soils with neutral soil reaction from formerly podzolized acid soils. Researchers in the field of soil genesis use the term "polygeneticity" to describe this phenomenon.

For a mature soil (third stage of soil development) any cycle of biologic (and a more universal-biogeochemical) cyclicity almost repeats any previous cycle. The involvement of new elements from the minerals of a parent material occurs within certain limits. The essential soil properties may appear relatively unchanged with the passage of time. At the stage of soil development the equilibrium among all types of pedogenic processes (micro-, meso,- and macroprocesses, biologic and abiotic ones) takes place in the soil profile. Any virgin soil reclamation and its involvement into agriculture disturbs the formerly acquired equilibrium a rational use of a soil should be based on the laws of soil formation and the predictions of soil development under the impact of practices involved in crop production systems, the use of pastures and grasslands, fruit orchards and other specialized modes of soil use. Methods of soil reclamation (drainage and irrigation, neutralization of excessive acidity or alkalinity, fertilizing systems, etc.) should increase reproduction of soil fertility and productivity in general. The consequential stages of soil development are shown by a hypothetical curve in Fig 3.1.

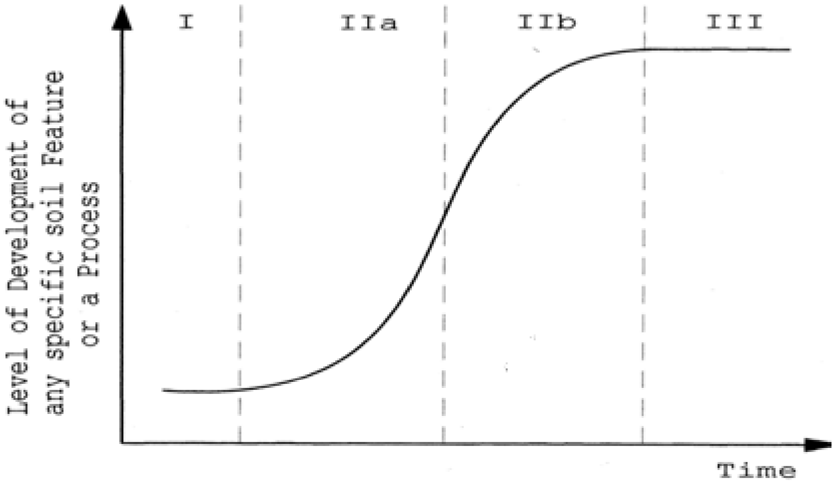


Fig 73. The curve characterizing the Stages of Soil Development: I – the initial stage of a soil-forming process;

IIa – the accelerated substage of soil development;

IIb – the slowing down substage of soil development;

III – the stage of soil maturity (climax).

The initial, or primary, stage of soil development is characterized by the absence of clearly detectable soil features in the solid phase. Microprocesses taking place in are not organized into meso- or macroprocesses. Biological cycle is weak, not involving large amount of matter.

The stage of accelerated soil development (IIa) is characterized by progressively increasing capacity of biological cycles. Every previous cycle improves the conditions for the next one. Biodiversity of soil organisms progressively increases. Mesoprocesses get their shape and intensity. Specific differentiation of a soil profile is becoming ever more evident, involving the aggregates of soil structure and soil horizons.

The stage of soil development diminishing in rate (IIb) is characterized by the lowering rates of a biological cycle. The intensity of the cycles increases by lesser and lesser increments of matter and energy involved. Profile-forming macroprocesses develop simultaneously with micro-and mesoprocesses of soil formation.

The stage of soil's mature functioning (III) has is characterized by the following features: 1) all essential soil-forming processes occur within the frames of relatively stable biogeochemical cycles; 2) there is a relative stabilization of essential soil features (morphological, etc), except the features subject to seasonal changes.

Energy and Soil Formation. The main and practically the only source of energy entering the soil is solar radiation. The entire surface of the earth receives from the sun approximately $21{\cdot}10^{20}$ kilojoules of energy annually. Most of this energy is consumed in the evaporation of water from the surface of land and oceans and in the turbulent heat exchange between the earth's surface and the atmosphere (that is on the formation of climate and oceanic currents). Photosynthesizing organisms (green plants) assimilate only about 0.5-5% of solar energy.

According to V.R.Volobuyev, the losses of energy for soil formation under natural conditions are determined mainly by the radiation balance, humidity coefficient, and biological activity of the biogeocaemosis. One of the scientist's formulas allows computing the discharge of energy for soil formation in calories per square centimeter per year (Q):

$$Q = R \cdot \exp(-\frac{18.8R^{0.73}}{P}) \tag{15}$$

where R is the radiation balance of a location in calories per square centimeter per year; and P is the annual precipitation in mm. In agricultural sector, the losses of energy for soil formation are also affected by the heat capacity of the soil, soil moisture regime, and the productivity of crops.

12.4. Brief Outline of Soil Formation Processes

Elementary pedogenic processes (particular macroprocesses of soil formation, according to I.P. Gerasimov, 1973) are specific only for soils and not for any other natural medium. They determine the formation of specific soil horizons and the constitution (construction) of the soil as a natural body. They may take place in a series of soil types in various combinations and with varying individual intensity. According to B.G. Rosanov's classification, they are united into the following groups:

1) biogenic-accumulative;

2) hydrogenic-accumulative (soil water, not hydrogen, determined origin);

3) metamorphic;

4) eluvial;

5) illuvial-accumulative;

6) pedoturbation, and

7) destructive.

Biogenic-accumulative elementary soil-forming processes take place under a direct influence of living organisms, the products of their metabolism, and their posthumous residues which form the surface horizons of soil organic matter accumulation. Some examples of such processes may be forest litter formation, peat formation, humus accumulation, sod-forming process.

Forest litter (forest floor) is an organic horizon (H_0) that developed primarily from the accumulation of leaves, twigs, and woody materials with or without minor components of mosses. Forest litter formation occurs only under natural conditions of virgin forest soil formation. The lower part of the forest litter is an organic-mineral subhorizon.

Peat formation occurs commonly at the surface of mineral soils. Organic horizons, according to the Canadian system of soil classification, should contain not less than 17% organic C (30% SOM) by weight. Soil development takes place under conditions of extreme wetness. Vegetation produced in the shallow waters of lakes and ponds may accumulate as sediments of peat because of a lack of oxygen for their decomposition. Organic soils have T horizons dominated by undercomposed organic material. T horizons are called histic horizons. The degree of decomposition of the organic matter in histic horizons is indicated by different symbols (H_t, TH, T_h, $T_{(h)}$, etc.).

Soil humus formation is the process of organic residue transformation into soil humus and its mixing with the mineral soil with formation of humus separates (humones) enveloping mineral particles and forming organic-mineral derivative compounds with silicate and nonsilicate clays.

Sod-forming process is an intensive soil humus formation and accumulation accompanied by nutrient element accumulation under the influence and from the residues of grass vegetation, especially its root systems, with the formation of isohumic soil profile with dark-colored and intensively granulated surface horizon (humus horizon) containing abundant grass roots (H_d horizon) and therefore compacted though well penetrable for roots and percolating water. About one half of the soil volume is occupied by the roots. The process is favored by the capillary fringe coming to the surface of the soil or at least to the roots of grass.

Hydrogenic-accumulative elementary processes of soil formation. Let us take for an example the development of soil salinity from saline ground water. Salt-affected soils are often found within irrigated lands. Salt-affected soils are common in arid and semiarid regions, where annual precipitation is insufficient to meet the evapotranspiration needs of plants. As a result, salts are not leached from the soil. From saline ground water the salts, by capillary rise, reach the soil horizons and the roots of plants. They accumulate in the surface horizons from plant residues. When capillary fringe reaches the surface, the evaporation of water may cause the formation of salt crusts on the surface of soils. As much as one-third of all irrigated lands in the world may be plagued by salt problems.

Accumulation of gypsum may be another example of hydrogenic-accumulative processes of soil formation, when sodium sulfate containing ground waters reach by capillary rise the calcium carbonate containing layer of soil profile (and $CaSO_4$ solubility product is attained):

$$Na_2SO_4 + Ca(HCO_3) = CaSO_4 \cdot 2H_2O + 2NaHCO_3$$

Calcium carbonate accumulation occurs with the rapid desiccation of the upper diffusive boundary of capillary fringe in accordance with the equation:

$$Ca^{2+} + CO_2 + H_2O = CaCO_3 + 2H^+$$

Soil desiccation in this case allows reaction $CaCO_3$ solubility product.

Metamorphic processes of soil formation are in their essence linked with the in situ transformation of rock-forming minerals without any eluvial-illuvial translocation of transformation products in the soil profile.

Siallinization is the process of primary mineral weathering in the soil with the accumulation of secondary silicate clay. Sometimes the process is termed *clay formation*. Mineral formation in soils is very often represented by the formation of some secondary 2:1 layer silicates like montmorillonite. In B.G. Rosanov's classification the process is known as montmorillonitization. The exact chemical conditions under which soil minerals form are not known at present.

Gley formation is another example of metamorphic transformation of the soil's mineral part under conditions of continuous or periodic excessive wetness leading to the development of reduction processes occasionally alternated by oxidation in local zones of soil materials. Water saturation occurs because of the location of the soil in a landscape position that is naturally poorly drained or where an impermeable layer causes saturation. Soil horizons or the entire profile become of dull grey, greenish grey, bluish grey, olive and other characteristic colors which is caused by the presence of Fe^{2+} and other reduced elements.

Eluviation processes of soil formation are linked with the destruction or transformation of a soil material in a specific eluvial horizon and translocation of the products by percolating or lateral water fluxes into lower horizons. Eluvial horizon becomes poorer in some compounds and richer in those that were left in situ.

Soil leaching is the least intensive of the eluviation processes. The soil horizons may be leached calcium carbonate or soluble salts. Eluvial horizons may be of different nature in podzolic, podzolozed, sodic, and solot soils.

Podzolization is important for the formation of podzols and podzolized soils. The podzol soils develop best of all in a cool, moist climate in coniferous or mixed coniferous and hardwood forests. The needles are highly resinous and decompose slowly. The forest trees, especially the pines and spruces, feed relatively lightly on bases and return less of these to the surface to counteract the leaching influence of the rainfall than do the grasses. The soil becomes acid, and the decomposition of the organic matter is accomplished more by fungi which favor acid conditions and less by bacteria than in grassland areas. Soil minerals are subject to acid hydrolysis. The humus formed has a fulvatic composition. Soluble products rich in R_2O_3 are leached down. The surface layers become grey and are strongly leached of both organic matter and plant nutrients. In the illuvial horizons some of the flocculated colloids and other insoluble materials accumulate.

Lessivage ("leaching" of clay particles) occurs when deflocculated clay particles are translocated from the upper to the lower horizons of soil profile. Under conditions of high precipitation, the loss by lessivage of clay particles without their previous destruction (like hydrolysis in podzolization) tends to leave behind a horizon of lighter color and poorer in clay content. Micromorphological inspection shows the platelets of silicate clay oriented in space parallel to the direction of water flow.

Podzolization in combination with the lessivage causes the formation of spodic, albic and other eluvial horizons.

In sand-textured parent material there is little clay for illuviation and rapid water permeability. Under conditions of high precipitation, iron and aluminum oxides (sesquioxides, R_2O_3) are released in weathering. These oxides are believed by I.S.Kaurichev to complex or chelate with organic matter (fulvic acids) and are transported to the subsoil. The loss by eluviation of sesquioxides and clay tends to leave behind a light-colored elluvial horizon called in the USA-system the albic horizon. The Americans derive it from the word white. The Russians see it to be light ashy-grey in color (podzolic).

Illuvial-accumulative elementary pedogenic processes correspond to the respective processes of eluviations and are their consequence. All illuvial processes result in the formation of illuvial horizons. An illuvial subsoil horizon that is greatly enriched in the amorphous compounds of R_2O_3 formed and transported in podzolization is called a **spodic horizon**. In the American system of symbols spodic horizons can B_e, B_S, B_h, or B_{hs} horizons depending on the degree of accumulation of sesquioxides, humus, or humus plus sesquioxides, respectively. In Ukraine symboll (illuvial) is employed to designate it.

In sodic soils containing over 15% of exchangeable Na^+ of the total CEC, this cation disperses soil colloids and causes their translocation to illuvial sodic horizons where they flocculate, again and form columnar or nutty structure. Illuvial horizons of sodic or sodicity-affected soils have alkaline (pH>8.5) reaction and are impenetrable to roots. In the American system of horizons classification such horizons are called nitric.

Any gradual illuvial accumulation of clay in a horizon converts it into argillic (B_t) horizon. **Argillic horizons** usually require several thousand of years to form (H.D. Foth, 1990). The processes of mechanical mixing of soil material under the impact of natural and anthropogenic agents are called **pedoturbation**. They include soil cracking in vertisols, self-mulching, cryoturbation and other processes.

Of the **destructive elementary processes of soil formation** it is worthwhile to mention soil erosion by water and wind, **soil downslope creep during plowing and other forms of soil tillage** and soil burial by the other materials turning it into a relic soil.

Generally speaking, the elementary processes of soil formation are very numerous, and the essence and mechanisms of many of them have not yet been deciphered.

12.5. Morphological Features of Soil Profile

The first stage of the field soil survey sets its a task to diagnose and classify soil units by the external features of a soil profile using simple equipment and reagents. A specialist in soil survey has to determine the soil type, subtype, genus, species, variety, and lithologic series. According to M.I. Polupan et al (2005), “morphological features of a soil are its external, ecologic-functional features using which it is possible to determine a certain genetic type this soil belongs to”. They include the soil’s constitution as a natural body (the set of soil horizons), soil and soil horizon depth, soil color, soil texture, soil structure, soil tilth, new formations, inclusions, the character of transition and the shape of a “borderline” between the horizons.

Each genetic horizon of the soil profile is designated by a letter symbol. The Ukrainian system of horizons symbols had been proposed by O.N.Sokolowsky and is now slightly modified and improved. The surface horizon of humus accumulation is designated by H (humus), eluvial horizon (eluo-wash, leach) – by E, illuvial (illuo-wash into) – by I, peat accumulation horizon – by T, peat and humus accumulation horizon – by TH, a horizon transitive between organic and mineral horizons – by TC, surface horizon of undercomposed residue accumulation – by H_0, the horizon of sod enriched with grassroots and grainy structure – by H_d, illuvial horizon of sodic (solonetz) soils – by Sl, gleyed (subject to reduction processes) horizon – by Gl, parent material – by P, material underlying parent material – by D. The system has an advantage of combining the symbols to designate the intermediate (transition) horizons of the profile and of using both capital and small letters depending on the extent of a process development (H_p, P_h, PE, P_e, etc). Small letters are also employed to designate new formations and inclusions (K – for carbonates, s – for soluble salts, n – for solid concretions, Kn – for carbonate concretions, q – for chert, z – for earth warm castings, etc.). **A vertical exposure of a soil consisting of horizons is a soil profile.** Soil constitution as a natural body may be expressed by the sum of horizons from the surface downward: H_0 + HE + E + I + P (soddy-podzolic soil), H + H_{plk} + P_{hk} + P_k (typical chernozem), etc.

Soil horizons are genetically related; however the layers differ from each other in their physical, chemical, and biological properties. Soil layers are approximately parallel to the land surface and several layers may evolve simultaneously over a period of time. The idea of designating soil horizons with A, B, and C symbols was V.V.Dokuchaev. By studying soils of varying age, soil scientists have reconstructed the kinds and the sequence of changes that occurred to produce soils. The uppermost horizons of organic matter accumulation in mineral parent materials are A horizons. The unaltered or only slightly altered upper part of the parent material is the -horizon. The soil at this stage of evolution has only two horizons – the A horizon and the underlying C-horizon. The subsoil layer eventually develops a distinctive color and some other properties that distinguish it from the A horizon and underlying parent material. This altered subsoil zone becomes a B-horizon. At this stage of soil evolution, there has not been enough time for the B-horizon to be significantly enriched with fine-sized (colloidal) particles. These particles may be translocated downward from the A horizon by percolating water. In the American system of horizons, such a weakly developed B-horizon is designation given the symbol w (weak), as in B_w. A B_w horizon can be distinguished from A and C-horizons primarily by color, arrangement of soil aggregates, and an intermediate content of organic matter. When a significant increase in the clay content of the B_w horizon occurs due to clay translocation, a B_w horizon becomes a B_w horizon. The t (as in B_t) refers to an illuvial accumulation of clay. Many sand parent materials contain very little clay. As a consequence, clay accumulation is insignificant and B_t horizons do not evolve. Humus, however, reacts with oxides of aluminum and /or iron to form complexes in the upper part of the soil. Where much water for leaching (percolation) is present, as in humid regimes, these complexes are translocated downward in percolating water to form illuvial accumulation of humus and oxides (R_2O_3) in the B-horizon produces B_{hs} horizons. The h indicates the presence of an illuvial accumulation of humus and the **S** indicates the presence of illuvial sesquioxides (R_2O_3). B_{hs} horizons are common in very sandy soils that are covering the so-called "borova" (bor = pine forest) terraces of the rivers in Ukraine. Ancient alluvial deposits are the parent materials of soils on such teraces. The soils have low fertility and low water-retention capacity (drought). In soils with thin A horizons, a light-colored horizon may develop at the boundary of the A and B horizons. This horizon, commonly grayish in color, is the E horizon. The symbol E is derived from eluviations, meaning "washed-out". Organic soils in American system of horizon symbols have O horizons; the O refers to soil layers dominated by organic material. In some causes, extreme wetness and acidity at the surface of the soil produce conditions unfavorable for decomposition of organic matter. The result is the formation of O horizons on the top of mineral soil horizons, Thus in present-day American soil morphology the master soil horizon or layers include the O, A, E, B, C and R horizons.

The depth of the solum (soil profile from the surface to parent material) is in our country measured in centimeters. The thickness of soil horizons is also measured in centimeters, for example:

$H_0\frac{0-3}{3}$; $HE\frac{3-18}{15}$; $E\frac{18-48}{30}$; $I\frac{48-98}{50}$; etc.

The upper and the lower borders of each horizon are stated as well as their absolute thickness in the denominator.

Soil color is one of the most important diagnostic features of soils. Many soil types received their name by their surface soil color, like chernozems, brown forest soils, podzolic (ashy-grey) soils, chestnut soils, greyzems, cinnamonic soils, etc. The most objective way to determine soil color is by using standard scales such as Munssel Soil Color Charts, in which soil color is compared to colored spots characterized by three characteristics: 1) dominant spectrum color, determined by the dominating wave length; 2) shade or the expressiveness of spectral color and 3) intensity – chromatic purity and extent of illumination linked with the amount of reflected light. Thus, index 10 YR 6/3 is deciphered like this: 10 YR – 10 yellow-red (chroma) , intensity:6, shade:3. Such an evaluation pertains to yellowish brown color. M.I. Polupan et al (2005) propose the following specter of colors to describe the profiles of Ukrainian soils:

Black	Dark Grayish Green
Dark grey	Light Grayish Green
Grey	Light–reddish Brown
Light Grey	Straw–yellow
Whitish Grey	Dark straw–yellow
Brownish Grey	Yellowish straw-yellow
Dark-brownish Grey	Brownish straw-yellow
Light-brownish Grey	Brownish – Grayish straw yellow
Brown	Yellow (canary)
Dark Brown	Dark yellow
Grayish Brown	Light yellow
Dark-grayish Brown	Light Brownish-yellow
Light- grayish Brown	Brownish–yellow
Darkly Brown	Dark Brownish – Yellow
Dark – yellowish Brown	Brownish Cinnamonic
Light – yellowish Brown	Dark Brown Cinnamonic
Reddish Brown	Light Brown Cinnamonic
Light Reddish Cinnamonic	Red-Cinnamonic
Reddish Cinnamonic	Dark Reddish Cinnamonic
Dark Reddish Cinnamonic	Dull Grey
Light Reddish Cinnamonic	Dark Dull Grey
Red	Light Dull Grey
Dark Red	Grayish Dull

Cinnamonic Red	Dullish Grey
Dark Cinnamonic Red	Sky Blue
Light Cinnamonic Red	Dark Sky Blue
Olive	Light Sky Blue
Light Olive	Dull- Grayish Sky Blue
Grayish Olive	Blue
Light Grayish Olive	Cinnamonic
Grayish Olive	Dark Cinnamonic
Olivish Grey	Light Cinnamonic
Dark Olivish Grey	Grayish Cinnamonic
Light Olivish Grey	Dark-Grayish Cinnamonic
Grayish Green	Light-Grayish Cinnamonic

S.A. Zakharov's Triangle of Soil Colors widely used in Ukraine since Soviet times admits three groups of substances determing soil color: 1) humus (black); 2) iron oxides and hydroxides (red), and 3) silica and carbonate of lime (white). The sides of the triangle, its median and the center represent the intermediate shades of color. But the triangle does not adjust itself to gleyed soils and horizons with their greenish, olive, bluish, and bright blue (vivianite) colors.

Soil Texture in the former (before Polupan et al, 2005) soil classification determined soil variety (type, subtype, genus, species, variety, lithologic series). In the up-to-date classification it determines soil *genus.* In the field one of the field methods is employed preferably a "moist variant" in which the soil sample is moistened to the "consistence of putty", and the ability of the soil to mold into "balls", "wires" and "threads" based on the plasticity index and described in the previous chapter, is evaluated.

Soil structure is determined in the field by way of tossing the soil up on a spade and examining the aggregates by their shape and size. The best classification of soil aggregates was proposed by S.A. Zakharov and is presented in the following table.

Table 24. Classification of Soil Structure

Genus	Species	Size
First Type: Cubic Structure		
Blocky: irregular shape and nonuniform surface	Large blocks Small blocks	> 10 cm 10-1 cm
Cloddy: irregular spherical shape and nonuniform spherical and rough (broken) fracture surfaces, faces not expressed	Large clods Clods Small clods Dust	10-3 mm 3-1 mm 1-0.25 mm < 0.25 mm
Nutty: more or less regular shape, faces well expressed, surface uniform, edges well delineated	Large nuts Nuts Small nuts	> 10 cm 10-7 mm 7-5 mm
Grainy: more or less regular shape, occasionally spherical with leveled faces, which may be alternatively rough and dull or smooth and with luster	Large grains (pea-like) Grains (cereal-like) Small grains (powder-like)	5-3 mm 3-1 mm 1-0.5 mm
Second Type: Prismatic Structure		
Column-like: the aggregates are not distinctly shaped with uneven faces and smoothed edges	Large column-like Column-like Small column-like	> 5 cm 3-5 cm < 3 cm
Columnar: of regular shape with well expressed smooth lateral and vertical faces with nearly spherical upper base (head-like) and flat lower base	Large columnar Small columnar	5-3 cm < 3 cm
Prismatic: with well expressed faces and smooth "varnished" surfaces and sharp edges	Large prisms Prisms Small prisms Fine prisms Pencil-like (with the length over 5 cm)	5-3 cm 3-1 cm 1-0.5 cm < 0.5 cm < 1 cm

Third Type: Platy Structure		
Plate-like (layered): with more or less developed horizontal planes of “cleavage”	Slate-like platy	> 5 cm
	Thin-platy	5-3 cm
Scale-like: with relatively small slightly warped horizontal platelets and very often sharp faces (looking distantly like fish scale)	Leaf-like	< 1 mm
	Shell-like	> 3 mm
	Rough-scaly	3-1 mm
	Fine-scaly	< 1 mm

To ”diagnose” a soil structure in the field it is necessary to use the drawing depicting S.A. Zakharov’s classification and complementing it like a practical visual aid. (Fig. 74).

Fig 74. Essential Species of Soil Structure:

I type: 1 - large clods; 2 - clods; 3 - small clods; 4 – dust; 5 - large nuts; 6 - nuts; 7 - small nuts; 8 - large grains; 9 - grains (crumbles); 10 - powder; 11 - “necklace” of soil grains; 12 - columnar aggregate; 13 - column-like aggregate; 14 - large prisms; 15 - prisms; 16 - small prisms; 17 - fine prisms; 18 - slate-like aggregates; 19 - platy aggregates; 20 - leaf-like aggregate; 21 - rough-scaly aggregates; 22 - fine-scaly aggregates.

More often than not the soil structure is represented by mixed genera and species, and even mixed types. Cloddy and grainy structure is characteristic of sod-forming horizons, platy and leafy – of eluvial horizons, nutty – of illuvial horizons of gray forest loamy soils, prismatic – of illuvial horizons of podzolic and soddy-podzolic soils, etc. Columnar structure with smoothed “heads” is typical for the illuvial (sodic) horizons of sodic (solonetzic) soils, containing considerable amount of exchangeable sodium.

Soil Tilth and Porosity Condition is (складення in Ukrainian) an important characteristic of a solid phase, determining soil density and penetrability to roots, water retention and infiltration, soil aeration, yielding to tillage operations, etc. This soil condition considerably depends on the way the soil particles and aggregates are arranged. By density and compaction soils are divided into the following groups:

1) very compacted: it is impossible to dig a pit with a spade; a pickaxe is needed for the purpose;
2) compacted: much effort is needed to dig a pit, but the task can be performed without a pickaxe;
3) loose: it is relatively easy to dig a pit with a spade; this can be done
4) spilling: like that of dry coarse-grained sands.

The visually identifiable condition of soil porosity may be the following:

1) finely porous soil: with pores less than 1 mm in diameter;
2) porous soil: pore diameter within 1-3 mm, like in loess;
3) spongy soil: as the name implies – numerous pores within 5-10 mm in diameter;
4) nostril-like porosity: soil holes of 5-10 mm in diameter;
5) honeycomb – like porosity: soil cavities of over 10 mm in diameter;
6) tube - like porosity: soil is rich in mole or other earth-moving animals and rodents tunnels.

Soil pores between the aggregates look like cracks. The following cracks are identified morphologically:

1) fine cracks: thinner than 3 mm;

2) cracks: 3-10 mm thick;

3) chinks: thicker than 10 mm.

New formations and inclusions. New formations are the accumulations of substances of varying form and chemical composition formed as a result of soil formation and deposited in soil horizons. New formations are roughly classified into those of chemical and those of biological origin.

Chemical new formations may be represented by the efflorescences of soluble salts or gypsum, accumulations of calcium (and magnesium) carbonates of various forms, spots of crystalline silica, precipitates of aluminum and iron oxides and hydroxides in the form of streaks, spots, and layers of varying thickness, concretions of iron and manganese as “peas”, “beans”, and “lentils”, and numerous other formations. New formations allow deciphering the essence and intensity of micro- and mesoprocesses of soil formation. Thus zonal subtypes of chernozem soils, formed in loess, whether forest-steppe or steppe pedons can at the very first glance be determined by the character of new formations of calcium carbonate in the so-called illuvial-cabonate horizons: steppe soils (ordinary and southern chernozems) have pea-like concretions of $CaCO_3$ in these horizons, whereas forest steppe (typical) chernozems may have carbonate “mold” or “veins”.

Biological new formations are represented by warm channels and casts (coprolithes), insect pellets (excrements), “crotovinas” (mole tunnels), cavities left by rotten roots, patterns of small roots left on the faces of soil aggregates, etc.

Inclusions are various “foreign” bodies of organic or mineral origin, the presence of which is not directly linked with the process of soil formation. Boulders and other rock fragments in soils formed from glacial till or from eluvial of solid rocks, bits of “ancient” bricks and porcelain, shells and bones of animals, etc.

The borderlines between the soil horizons and the character of transition from the upper to the lower horizon also are the object of soil morphology. The transition may be sharp or distinct, if the changes of properties occur within 3 cm, clear (3-5 cm), and gradual (> 5 cm). The borderlines may be direct, wavy, tongue-like, pocket-like, etc. Owing to the imperfection of our national soi; taxonomy, some of the terms used in a textbook may be not sufficiently lucid for the readers from other countries. Much work still remains to be done.

Questions & Assignments

1. Name and characterize the natural factors of soil formation.
2. Which five “universal” processes compose the general scheme of soil formation?
3. Name and briefly describe the mesoprocesses of soil formation known to you.
4. What do you know about the morphologic features of a soil profile and for what purpose are they used?
5. Describe soil as an ecological system.

Chapter 13

Organisms and their role in soil formation and development of soil productivity

Three groups of organisms take part in soil formation: green plants, microorganisms, and animals which form terrestrial associations – biocaenoses. Life in the soil is amazingly diverse, ranging from the smallest unicellular living things to large burrowing animals. The essential chains of soil formation – synthesis and destruction of organic matter, selective concentration of biologically important elements, disintegration and decomposition of rocks and minerals, movement and translocation of soil formation products – all these processes are the essential circumstances for the development of soil fertility and productivity. Soil organisms form an ecosystem which has its own unique combination of living organisms and abiotic resources that function to maintain a continuous flow of energy and nutrients. Soil, like any ecosystem, has two types of organisms based on carbon source: 1) producers, and 2) consumers and decomposers. Producers fix inorganic carbon from carbon dioxide and are autotrophs. Consumers and decomposers use the carbon fixed by producers, and are heterotrophs.

13.1. Green Plants

Vascular plants are the major producers that use solar energy to fix carbon from carbon dioxide during photosynthesis. About 5.3×10^{10} tons of biomass are synthesized by green plants on the terrestrial part of the globe annually. The tops of plants provide food for animals above the soil-atmosphere interface. Plants produce roots, tubers, and other underground organs within the soil that serve as food for soil-dwelling organisms. The main function of green plants as soil builders is sustaining by them of the biological cycle of substances and energy. The result of this process is the accumulation of potential energy and plant nutrition elements in the upper part of a soil profile. It is stimulate a gradual development of soil fertility and productivity as a whole (Table 25).

Table 25. Biological Cycle of Substances (mt/ha) according to L.E.Rodin and N.I.Basilevich (1965)

Plant Association	Organic Matter				
	Total Biomass	Including Root Biomass	Annual Increment	Annual Dying Out	Organic matter in Forest Litter and Steppe “Blanket”
Arctic Tundras	5.0	3.5	1.0	1.0	3.5
Pine Forest of Southern Taiga	280	63.6	6.1	4.7	44.8
Spruce Forest	330	73.5	8.5	5.5	35.0
Birch Forest	220	50.5	12.0	7.0	30.0
Sphagnum Forest Bogs	37.0	4.0	3.4	2.5	> 100
Forest Dominated by Oaks	400	96.0	9.0	6.5	15
Meadow Steppes	25.0	17.0	13.7	13.7	12
Arid Steppes	10.0	8.5	4.2	4.2	1.5
Semibush Deserts	4.3	3.8	1.2	1.2	–
Dry Savannahs	26.8	11.3	7.3	7.2	–
Wet Subtropical Forests	410	82.0	24.5	21.0	100

The ABC of understanding of plant association role in soil formation and productivity development were proposed by V.R. Williams, a noted Soviet agronomist and academician. V.R. Williams proposed to differentiate the following plant associations important for the soil survey and generalization.

1) associations (formations, according to V.R. Williams) dominated by tree plants: taiga forests, deciduous forests, dry and humid subtropical and rainy tropical forests;

2) transitory (intermediate) association of trees and grasses: forest-steppes, savannahs, etc.

3) association of grasses: dry-land and swampy meadows, prairies, moderate climate steppes, subtropical and bush steppes.

4) desert plants associations: subboreal with summer cycle of plant growth, subtropical with winter cycle of plant growth, and tropical;

5) associations of mosses and lichens (tundras and swamps).

Table 26. Cycle of Nitrogen and Ashy Elements in Different Plant Association (%)

Plant Association	Nitrogen and "Ashy" Elements				Nitrogen			
	In biomass	consumed annually	Returned with residues annually	in forest litter and Steppe "blanket"	in biomass	consumed annually	returned annually with residues	% of total chemical elements in residues
Arctic Tundras	0.16	0.038	0.037	0.28	0.081	0.021	0.020	53
Pine Forest of Southern Taiga	1.88	0.085	0.058	1.73	0.664	0.027	0.016	28
Spruce Forest	2.70	0.155	0.120	1.30	0.720	0.041	0.035	29
Birch Forest	2.10	0.380	0.290	1.60	0.875	0.150	0.090	30-40
Sphagnum Forest Bogs	0.61	0.109	0.073	–	0.229	0.040	0.025	–

Forest Dominated by Oaks	5.80	0.340	0.255	0.80	1.150	0.095	0.057	19-26
Meadow Steppes	1.18	0.682	0.682	0.80	0.274	0.161	0.161	22-28
Arid Steppes	0.35	0.161	0.161	0.07	0.103	0.045	0.045	17-36
Semibush Deserts	0.185	0.059	0.059	–	0.061	0.018	0.018	24-31
Dry Savannahs	0.978	0.319	0.312	–	0.238	0.081	0.080	26
Wet Subtropical Forests	5.28	0.993	0.795	0.60	1.359	0.277	0.226	28

The explanation for the differences in the amount and distribution of organic matter is related to the differences in the growth of the two kinds of plants-trees and grasses. The roots of grasses are short-lived, and each year they contribute to the soil large amounts of organic matter that becomes humified. There is a certain correlation between a gradual decrease in root density, an increase in soil depth and a gradual decrease in organic matter content. In the forest, by contrast, roots are long-lived and the annual addition of plant residues is largely as leaves and dead wood that fall directly onto the soil surface. Small animals transport and mix some of the organic matter with a relatively thin layer of topsoil. The differences in the uptake of ions and, consequently, in chemical composition of plants play a role in soil development. Species that normally absorb large amounts of the cations like calcium, magnesium, potassium, and sodium delay the development of soil acidity because they recycle more of these cations to the soil surface through the addition of O.M. The effect't delays their removal from the cation exchange sites and delay the development of acidity and the accumulation of exchangeable aluminum and hydrogen. Thus hardwoods maintain a higher pH than do spruce trees when grown on parent material with the same mineralogical composition.

Forest soils are more acid then grassland soils, due to greater leaching of basic cations. The study of the effects of trees versus grass on soil genesis shows that the same fundamental processes occur in both kinds of soil but differ in degree of expression. Root systems of grasses and the products of their humification granulate the upper, rich in roots, portions of soil profiles.

Farmland vegetation of agricultural crops grown in crop rotations in Ukraine can supply the soils with 3-8 mt/ha of crop residues (surface residues + roots). Some 10 to 30% of this amount turns into soil humus and the remainder is mineralized to CO_2, H_2O, NH_4^+ and other mineral compounds.

13.2. Soil Microorganisms

Soil is the home of bacteria, fungi, actinomycetes and algae composing the associations of forest, meadow and other biocaenoses. Microorganisms are both consumers and decomposers. They are consumers in the sense that they consume materials for growth and, at the same time, they are decomposers in the sense that carbon is released as CO_2. A gram of soil may contain millions to billions of microorganisms. (Table 27).

Table 27. Number of Microorganisms in Eurasia Soils (data of E.N. Mishustin)

Soil	Number of microorganisms, mln	
	Per 1 g of soil	Per 1 mg of soil N
Podzols (virgin)	300-600	about 70
Soddy-podzolic: virgin cultivated	600-1000 1000-2000	about 200 about 250
Chernozems (mollisols): virgin cultivated	2000-2500 2500-3000	about 600 about 750
Grey arid subtropic soils (sierozems): virgin cultivated	1200-1600 1800-3000	about 2000 about 2400

The greatest numbers of microorganisms are concentrated in the surface horizons. With depth their number drastically diminishes (Table 28).

Table 28. Distribution of Bacteria over Soil Horizons of a Soddy-Podzolic Soil (mln×g^{-1}, S.Rasumov and N.Remesov)

Horizon	Depth, cm	Meadow	Depth, cm	Plowland	Depth, cm	Forest
$A_1(H)$	1-4.5	955.3	2-6	1086	0-2.5	2693.0
	4.5-7.5	852.9	10-18	982.4	2.5-5	2246.6
	7.5-11	565.9	–	–	5-8	1781.5
	11-15	402.6	–	–	9-12	782.6
$A_1A_2(HE)$	15-19	87.1	16-18	618.2	12-15	517.0
$A_2(E)$	19-23	71.0	20-22	382.5	16-20	355.9
$A_2B_1(EI)$	23-28	50.8	–	–	21-25	265.6

Maximum quantity of microorganisms is located near living rootlets and on the surface of dead plant residues. They form a film rich in microorganisms called rhizosphere. But soil environment properties are changed by management practices. Biological processes are influenced by soil tillage and residue management. Where crop residues are retained on the soil surface (zero tillage) and crops are planted directly into a chemically-killed cover crop with a high C/N ratio, immobilization of soluble N can be much greater in no-tillage than plowed surface soils (Rice and Smith, 1984). The stratification of crop residues, organic matter, and soil organisms with no-tillage management is different than with conventional plowing (Table 29).

Increased microbial biomass is also often associated with higher density and activity of plant roots in surface no-tillage soils (Lynch and Punting, 1980). All forms of dead organic materials are attacked by decomposers, mainly by bacteria and fungi. Through enzymatic digestion the carbon is returned to the atmosphere as CO_2 (influencing hot-house effect) and energy is released as heat. The nutrients in organic matter are mineralized and appear as the original ions that were previously absorbed by plant roots. In this context it is appropriate to consider soil as **the stomach of the earth** (H.D.Foth, 1990).

Table 29. Comparison of Soil Properties and Microbial Parameters for No-tillage Relative to Plowing Across Six U.S.A. Locations (Doran, 1987)

Soil Depth, cm	Ratio, No-tillage versus Plow				
	Soil Water	Total Organic C	Total N	Mineralizable N	Microbial Biomass
0–7.5	1.3	1.4	1.3	1.4	1.5
7.5–15	1.1	1.0	1.0	1.0	1.0
15–30	1.1	0.9	1.0	0.9	1.0

The microorganisms are considered to be the major or ultimate decomposers. The major distinguishing feature of microorganisms is their relatively simple biological cellular organization. Many are unicellular and even multicellular organisms lacking differentiation into cell types and tissue characteristic of plants and animals. They form spores, or a resting stage, under unfavorable conditions and germinate and grow again when conditions become favorable. For the majority of microorganisms, optimal temperature is within the temperature range of 25-35°C and soil moisture about 60% of full capacity.

Bacteria are the most numerous and widespread group of soil microorganisms. Bacteria belong to the smallest living organisms and can be characterized as single celled exceeding all other soil organisms not only in numbers but in kinds. A gram of fertile soil commonly contains 1017 to 1010 bacteria. They multiply rapidly by elongating and dividing into two parts. The numbers of bacteria are very variable being too sensitive to soil conditions. They fluctuate markedly not only with season but with two-three-day cycles. Researchers estimated that the live weight of bacteria in soils may exceed 2.000 kilograms per hectare. Soil bacteria are commonly classified under two heads: autotrophic and heterotrophic. Most of soil bacteria are heterotrophic. Their energy and carbon both come directly from the soil organic matter. The autotrophic bacteria obtain their energy from the oxidation of mineral constituents and most of their carbon – from carbon dioxide.

The general-purpose decay and ammonifying bacteria, as well as the fungi and actinomycetes are all heterotrophic in character.

Most of the bacteria are aerobes and require a supply of oxygen (O_2) in the soil atmosphere. Some bacteria are anaerobic and can thrive only in absence of oxygen in the soil atmosphere. Some bacteria, however, are facultative aerobes. They thrive well in an aerobic environment but they can adapt to an anaerobic environment.

Among the aerobic heterotrophic bacteria occurring in soil environments, are spore-forming Bac. mycoides and Bac. subtilis, and among those not forming spores – Pseudomonas fluorescents and Pseudomonas pyacyanea.

Anaerobic bacteria cause the processes of decay of the components of plant and microbial cells to simpler but underoxidized organic and subsequently mineral compounds. Very important for the formation of soil fertility is ammonification process (decomposition of the proteins to ammonium):

Protein → aminoacids → NH3 + nitrogen-free organic compounds. Interaction between microbial activity and mineralization of soil organic N are often controlled by environmental factors. Ammonification is associated with both aerobic and anaerobic heterotrophs.

Widely encountered in the soil are the processes of butanic acid fermentation of hydrocarbons (cellulose, hemicelluloses, pectin, etc), which proceed according to the equation:

$$C_6H_{12}O_6 = CH_3 - CH_2 - CH_2 - COOH + 2CO_2 + 2H_2 + Energy$$

This process is performed by the various species of Clostridium.

Biological denitrification (nitric respiration) proceeds in accordance with the equation:

$$5C_6H_{12}O_6 + 24KNO_3 = 24KHCO_3 + 6CO_2 + 12N_2 + 18H_2O + 1722 \quad Kjoules \quad of \quad energy$$

Pseudomonas fluorescents and Micrococcus denitrificans are the most active denitrificators in the soil. In soils with poor aeration this process leads to the losses of nitrogen, as many of the denitrification products are gaseous and escape from the soil. Just as it is natural for nitrogen to be lost from the soil by denitrofication. Nitrate–nitrogen in poorly drained soil is especially subject to denitrofication. Anaerobic conditions may exist in well-drained soils after rains, when the interiors of soil aggregates are wet. Normally, denitrification is detrimental to agriculture, because N is lost from soils. It helps to prevent an excess of nitrate in the ground water of irrigated lands where high rates of N -fertilizer have been used. Nitrate in deep ground water (below the biologically active zone) is not denitrified.

Desulfofication (microbial reduction of sulfates) is caused by anaerobic bacteria of the Desulfovibrio genus:

$$H_2SO_4 - H_2SO_3 - H_2SO_2 - H_2SO - H_2S \uparrow$$

Sulfur is lost from anaerobic soils in the above way. But it may also be retained in peats and other oxygen-lacking soils, forming FeS and FeS_2.

Nitrification is the biological oxidation of ammonium $(NH_4)^+$ to nitrite (NO_2^-) and to nitrate (NO_3^-) as a two-step process. Nitrification are involved by specific autotrophic bacteria. The reactions and organisms are as follows:

$$NH_4 + 1.5O_2 (Nitrosomonas) = NO_2^- + 2H^+ + 2H_2O + Energy$$

$$NO_2^- + 0.5O_2 (Nitrobacter) = NO_3^- + Energy$$

The nitrifying bacteria obtain energy from the above oxidizing reactions. The first step produces H^+, which contributes to soil acidity. Nitrification changes the form of available N from a cation to an anion. This is of no great concern in plant nutrition, since most plants readily use both forms.

Nitrification is affected by oxygen supply and soil pH. Nitrification is inhibited in water-saturated soils. When soils have a pH of 6 or more and are well aerated, there is an abundance of nitrifying bacteria that quickly oxidize NH_4^+ absorbed by plants and microorganisms. But NO_3^- is subject to leaching, whereas, NH_4^+ is absorbed onto cation exchange sites and resists leaching. The nitrification inhibitor is used to maintain the nitrogen in the ammonium form. According to A.N. Lebedyantsev, up to 300 kg/ha of $N - NO_3$ may accumulate in the soil in the course of a year when soil conditions favor nitrification.

Sulfofication (oxidation of H_2S, S, and $R - SH$ to sulfuric acid) is performed in the soil by the genera of Beggiatoa, Thiothrix, Thiophoga, etc, as well as by thionic bacteria of the Thiobacillus genus in accordance with the equations:

$$2H_2S + O_2 = 2H_2O + S_2 + Energy$$

$$S_2 + 3O_2 + 2H_2O = 2H_2SO_4 + Energy$$

The energy evolved is utilized by the bacteria for the assimilation of carbon from CO_2.

Iron (Fe^{2+}) **oxidation** is performed by the iron bacteria occurring in swampy soils by the equation:

$$4FeCO_3 + 6H_2O + O_2 = 4Fe(OH)_3 + 4CO_2 + Energy$$

Thread-like forms of bacteria (Gallionella and Leptothrix) are the most widespread in anaerobic soil environment.

Nitrogen fixation. The atmosphere is made up of 79% nitrogen, by volume, as inert N_2 gas. Dinitrogen fixation is the conversion of N_2 to ammonia and subsequently into organic forms utilizable in biological processes. Most N fixation is biological, being either symbiotic or nonsymboitic. N-fixing organisms contain an enzyme, nitrogenase, which combines with a dinitrogen molecule (N_2) and fixation occurs in a series of steps that reduces N_2 to NH_3. Molybdenum and cobalt microelements are required by the nitrogen fixing organisms.

Symbiotic legume fixation. Legume plants are dicots that form a symbiotic nitrogen fixation relationship with heterotrophic bacteria of the genus Rhizobium. A root hair in contact with the bacterial cell curves around the cell. The dormant bacterial cell then germinates and forms an infection thread that penetrates the root. Once inside the root, bacteria multiply and are transformed into swollen, irregular-shaped bodies called bacteroids. An enlargement of root occurs and eventually, a gall or nodule is formed.

In actual farming practice, the amount of nitrogen added to soils by the growth of legume crops depends on the amount of the crop that is harvested and removed from the soil. If the entire crop of legumes is plowed under, all the fixed nitrogen is added to the soil. Even though the production of a legume crop may not increase total soil nitrogen, the decomposition of the plant residues releases considerable available nitrogen during the following months.

Many **non-legume** plant species have root nodules and also fix nitrogen symbiotically. Red older is an example.

Nonsymbiotic nitrogen fixation. Certain groups of bacteria, living independently in the soil, can fix nitrogen. These bacteria fix nitrogen nonsymbiotically, and a dozen or more have been identified. The two organisms that have been studied best belong to the genus Azotobacter and the genus Clostridium. Azotobacters are widely distributed in nature. They have been found in soils with pH 6 or above. The greatest limiting factor affecting their distribution in soil appears to pH. These organisms are favored by good aeration, abundant organic matter, and suitable moisture. Clostridium is also a heterotrophic bacteria, but is an anaerobe that can thrive in soils too acid for Azotobacter. Both of these heterotrophs depend on decomposable organic matter for their energy supply.

Actinomycetes refers to a group of bacteria with a superficial resemblance to fungi. The actinomycetes resemble bacteria in that they have a very simple cell structure and are about the same size in cross-section. They produce a blanched filamentous network. But it is less extensive than in fungi. Many of these organisms reproduce by spores.

Actinomycetes are in great abundance in soils. They can vary in numbers from 1 to 36 million per gram of soil. It is generally believed that they are not as important as bacteria and fungi as decomposers. They are much less competitive when fresh portions of organic matter are added to soils. Only when very resistant materials remain do actynomicetes have good competitive ability.

The surface of the soil is the interface between the lithosphere and the atmosphere. At or near this interface, the quantity of living matter is greater than at any region above or below. As a consequence the A horizon contains more organic debris or food sources than the B and C horizons do. Although other factors besides food supply influence activity and numbers of microorganisms, the greatest abundance of decomposers typically occurs in the A horizons.

Fungi are heterotrophs that vary greatly in size and structure. Nearly 700 of them have been identified representing over 160 genera. Fungi typically grow or germinate from spores and their vegetative forms are of filamentous nature, or of a thread-like structure, called mycelium. Their threads may be either simple and restricted or profusely branched. Fungi readily extend their tissue and penetrate into the surrounding environment. The mycelium is the working structure that absorbs nutrients, continues to grow, and eventually produces reproductive structures that contain spores. For convenience of discussion the fungi are divided into three groups: 1) yeasts; 2) molds, and 3) mushrooms.

A gram of soil commonly contains 10 to 100 meters of mycelia fragment. Thus, the live weight of fungal tissue in soils is about equal to that of bacteria. Molds and mushrooms are the most common fungi in soil environment. Molds play a role infinitely more important than the mushroom fungi. They will develop vigorously in acid, neutral, and alkaline soils but they are mostly favored by a lowered pH. They are noticeably abundant in acid soils and are especially important in decomposing the organic residues of acid forest soils. They are most versatile in their ability to decompose organic residues. Some types of mushrooms are collected for food.

Cellulose, starch, lignin, as well as the more easily affected proteins easily succumb to the attacks of fungi. The woody tissues of the forest floor (forest litter) provide an abundance of food for certain fungi that are effective decomposers of lignin.

Between numerous varieties of mushrooms and the roots of higher plants an interesting and economically important association exists. The mycelia of the fungus infest the plant roots giving an association called **mycorrhizae**, a term meaning “fungus root”. There are two types of hyphae or micorrhiza: ectotrophic and endotrophic. The ectotrofic hyphae outside a root results in the formation of a sheath that surrounds the root.

Endomycorrhiza are more abundant than ectomycorrhiza. They benefit most of the field and vegetable crop plants. Hyphae invade roots and ramify both between and within cells, usually avoiding the central core of the roots. The host plants provide the fungi with food.

The benefits for the host plant include: 1) increased surface area of roots good for the absorption of water and nutrients, particularly phosphorus, 2) increased drought and heat resistance, and 3) reduced infection by disease organisms.

According to recent, **vesicular-arbuscular mycorrhizae** (VAM) fungi and the mutualistic associations they form with plants are also subsystems which perform a unique and important functional role in the rhizosphere of most agroecosystems. Future research needs to address the overall contribution of VAM to the rates, amounts, and forms of carbon, nitrogen, and phosphorus cycling through the rhizosphere with particular attention to soil productivity along with plant productivity in the long-term perspective.

Soil algae are chlorophyll-bearing organisms and must live on or very near to the surface of the soil. However certain forms seem to obtain their energy largely from organic matter and may readily exist within or below the surface horizon. Soil algae are divided into three general groups: 1) blue-green, 2) green, and 3) diatoms.

In subsoils most algae are present as resting spores, or cysts. Grassland seems especially favorable for the blue-green forms. In old gardens diatoms are often numerous. All of the ordinary types of algae are monerans. Algae participate in the weathering of rocks and primary in soil formation.

Lichens are very viable and resistant to growth conditions organisms. They can live in the hot dry desert as well as the cold, wet Arctic. This organisms can also survive on bare rocks, wooden poles, the sides of trees, and even on tops of mountains. The nature of lichens is of a dual character. A lichen is made up of a fungus and alga that live together. An alga, as a simple autotroph, uses sunlight to produce its food. Combined, the two organisms (alga and fungus) in lichen can live in many places that neither could survive if alone. The fungus part of the lichen provides the alga with water and minerals that the fungus absorbs from whatever the lichen grows on. The alga part of the lichen uses the minerals and water to make food for the fungus and itself. It is one of the best examples of symbiosis. Lichens are sometimes known as pioneers. They are often one of the first living things to appear in rocky barren areas. Lichens release acids that break down rock and cause it to crack. Dust and dead lichens fill the cracks, which **eventually become more fertile places** for the other organisms to grow (primary stage of soil formation). In time, the rocky area (say, in Carpathians) may become a lush green forest.

Microorganisms actively participate in destruction and formation of minerals. They mobilize the elements composing minerals, like iron, manganese, sulfur, calcium, phosphorus, aluminum. Enzymatic oxidation and reduction of minerals are important in this respect. Among others, T.V. Aristovskaya contributed much to the study the activity of certain genera, like Pedomicrobium, Micromonospora, Metallogenium, etc. These microbes are capable to utilize the organic-mineral derivative compounds containing iron and manganese. Microbes concentrate these elements in their cell walls and create thereby the new formations rich in iron and manganese in the form of concretions. **The intensity of microbiological processes in soils are very much effected by the practices of soil management.** The soil ecosystem represents an intricate balance between living and non-living components. Biological responses to changes in physical and chemical environments predominantly relate to biological needs for energy, an energy sink (usually oxygen), water, nutrients, suitable temperature, and living space, and the absence of harmful conditions.

Agricultural production systems involving tillage, residue and crop management and fertilizing practices, both directly, and indirectly, affect soil environmental factors that control the growth and activity of plants and microorganisms. The definition of primary soil properties which determine the soil biological activity and the soil environmental for plant growth and microbial activity requires establishment of interpretive guidelines for identifying how variations in soil type, climate, and cropping systems influence the importance and interaction of soil properties in controlling and limiting biological activity.

13.3. Soil Animals and Their Role in Soil-Forming Processes

Soil fauna are numerous and diverse. Soil animals can be considered both consumers and decomposers because consume organic matter and some decomposition occurs in the digestive tract. The largest soil animals (**soil macroanimals**) are chiefly: 1) rodents and insectivore, insects, 3) millipedes, 4) sow-bugs (wood lice), 5) mites, 6) slugs and snails, 7) centipedes, 8) spiders, and 9) earthworms (last but, by far, not least).

Soil animals, however, play a minor consumer-decomposer role in organic matter decomposition. Some of them are parasitic vegetarians that feed on roots, whereas others are carnivores that prey on each other.

Soil macroanimals, besides nematodes and thread worms, include ptrotozoa (probably the simplest forms of animal life). Protozoa are one-celled organisms but they are considerably larger than bacteria and of a distinctly higher organization. Some are merely masses of naked protoplasm (amoeba).

Soil protozoa are divided into the following groups: 1) amoeba, 2) infusoria, and 3) flagellates. The nutrient habits of protozoa are not sufficiently cleared up. Most of these depend on non-living organic matter as a source of food. They are heterotrophs and feed on bacteria, algae, and fungal spores. Being aerobic organisms protozoa are found mainly in the surface soil horizons. During droughts their numbers drastically reduce. "Feeding on soil bacteria they "rejuvenate" microbial associations which, as a result of it, become more biochemically active" *(Soil Science Textbook* edited by I.S. Kaurichev, 1989).

Worms are represented in soil environments by two important kinds. Microscopic roundworms (nematodes) are very abundant in soil animals. They are of economic importance because they are parasites that invade living roots. The other important worm is the ordinary earthworm.

Earthworms are perhaps the best known of the largest soil animals. "The Lumbricus terrestris, a reddish organism, and Allolobophora caliginosa, pale pinkish in color, are very common both in Europe and in eastern and central United States. Lumbricus terrestris was imported into the United States from Europe" (H.D.Foth, 1990). This worm makes a shallow burrow and forages on plant materials at night. Some of the plant material is dragged into the burrow where it is moistened and readily eaten. Other kinds of earthworms exist by ingesting organic mater found in the soil. Earthworms eat their way through the soil by ingesting the soil en masse. **Excreted materials (earthworm castings) are deposited both on and within the soil.** Ch. Darwin was the first naturalist to observe the earthworm activity with the purpose of measuring is quantitatively. He wrote: “Earthworms played a much more important role in the history of the world than it could be supposud. For their small bodies they possess great muscular power. A weight of no less than ten tons of dry earth annually passes through their bodies and comes to the surface of each acre of land”.

The mixing of soil materials in the digestive track of earthworms, the creation of channels, and production of castings, alter soil structure and leave the soil more porous. Channels left open at the soil surface greatly increase water infiltration. Earthworm casts on a cultivated field are estimated by some authors to weigh as much as 16 mt/ha. Compared with the soil itself, the casts are definitely richer in organic matter, total and nitrate nitrogen, exchangeable Ca^{2+} and Mg^{2+}, available phosphorus and the percentage base saturation. The worms also bring about a notable transportation of the lower soil to the surface. They increase the stability of soil aggregates, especially in virgin soils or under the mulches. They prefer a moist habitat, and therefore, live mostly in medium to heavy soils where moisture capacity is high, rather than in those of a sandy and droughty nature. The nutrition of some earthworms depends on certain lime-secreting glands. Perhaps because of this they are not usually found abundantly in soils that are low in replaceable calcium. For gardeners and those who wish to maintain a high earthworm population in their soils, surface protection is as essential as high organic matter, suitable moisture, good drainage and aeration, and low acidity.

Nematodes (roundworms) are microscopic worms and are the most abundant living organisms in soils. Under a 10-power hand lens they appear as tiny transparent threadlike worms. They lose their water readily through their skin. Later, they reactivate when conditions become favorable. Based on food preferences, nematodes can be grouped into those that feed on: 1) dead and decaying organic matter; 2) living roots, and 3) other living organisms as predators. Parasitic nematodes attack a wide variety of plants.

Arthropodes have an exoskeleton and jointed legs. Most have a kind of heart and blood system, and usually a developed nervous system. The most abundant arthropods are **mites** and **springtails**.

Springtails are primitive insects less than 1 mm long. They are distributed worldwide in big numbers. Springtails have a spring-like appendage under their posterior which permits them to flit or prong in every direction. They live in the macropores of the litter layers and feed largely on dead plant and animal tissue, feces (dung), humus, and fungal mycelia. They are restricted to moist layers.

Springtails are in the order Collembola, and they are commonly called collembola. Their enemies include mites and small beetles, centipedes, and small spiders.

Mites are the most abundant air-breathing soil living organisms. They commonly have a saclike body with protruding appendages and are related to spiders. They are small like Collembola. Some mites are vegetarians and some are carnivores. Most feed on dead organic debris of all kinds. Some are predaceous and feed on nematodes, insect eggs, and other small animals, including spring tails. Activities of mites include the breakup and decomposition of organic material, movement of organic matter to deeper soil layers, and maintenance of pore spaces (runways).

Millipedes and centipedes are elongate, fairly large soil living organisms, with many pairs of legs. They are common in forests, and overturning almost any log or stone will send them running for cover. Millipedes have many pairs of legs and are mainly vegetarians. They feed mostly on dead organic matter. Centipedes typically have fewer pairs of legs than millipedes, and are mainly carnivorous consumers.

Ants and termites. Their activities may be more important than those of earthworms. Ants transport large quantities of material from within the soil, depositing it on the surface.

Insects also shred plant residues to smaller bits which decompose, humify, and mineralize more readily.

Rodents, including mice, ground squirrels, marmots, gophers, and prairie dogs also inhabit the soil. Moles and other animals leave small and not too small mounds on the surface and tunnels filled with soil or parent rock materials called "crotovinas" (from the Russian crot = mole). Crotovinas are common in grassland, but not only. Sometimes the upper 2 to 3 meters of the soil formed from loess is mottled from the crotovina spots. So the rodents, like the earthworms, ants, and termites move large quantities of soil and may greatly alter the nature of soil horizons.

Questions & Assignments

1. Why is the soil an ecosystem?
2. What is the role of green plants in soil formation and soil productivity development?
3. Which groups of plant life associations were identified by V.R. Williams as important to soil survey?
4. Which are the functions of microbial decomposers in the development of soil fertility? Describe the roles of fungi and mycorrhiza in soil processes.

Chapter 14

Soil Organic Matter

Soil organic matter (SOM) consists of organic residues and humus. Organic residues of higher green plants, microorganisms and animals inhabiting the soil are a row material for the formation of soil humus. Soil organic matter and the processes of its transformation play considerable role in the formation of soil and its essential features and properties, including relatively stable soil fertility capable of its own reproduction or renewal, different forms of soil buffer capacity, sanitation-related soil properties, various forms of soil adsorption and exchange capacity, etc. SOM actively participates in plant nutrition, formation of agronomically favorable soil physical properties, migration of elements within the soil and the entire ecosphere.

14.1. Sources of Soil Organic Matter and its Composition

The organic matter in an ecosystem consists of the organic matter above and below the soil surface. Thus, the distribution of organic matter in a pine forest ecosystem, according to H.D. Foth (1990), expressed as organic carbon, was 38 percent in the trees and ground cover and 9 percent in the forest floor. The remaining 53 percent of the organic matter was in the soil and included the roots plus the organic matter associated with soil particles. In grassland ecosystems, much more of the organic matter is in the soil and much less occurs in the standing plants and grassland floor. Over 95 percent of the total organic matter may be in the soil, where grasses are the dominant vegetation.

The primary, or original, source of SOM in the production of the primary producers is the higher plants. They give 60-70 percent of the material needed for the formation of soil humus. Microbial population supplies nearly one-third of this material and the contribution of soil animals is practically negligible (0.1-0.2 mt/ha), but animal remains and wastes are richer in nitrogen, and they produce more humus per unit weight. While in the tundra zone the annual input of organic residues to the soil is within 1-2 mt/ha (air-dry weight), in meadow steppes their amount is about 13-15 mt/ha. In agroecosystems the annual input of residues (surface plus root residues) may be as small as 2-3 mt/ha (intertilled or row crops) and as large as 7-9 mt/ha (perennial grasses). This organic material is subsequently consumed and decomposed by soil microorganisms. The result is its decomposition, mineralization and accumulation in soils that have great diversity and highly variable composition. The plant residues, including tops and roots, contain a wide variety of compounds: sugar, cellulose, hemicelluloses, protein, lignin, fat, wax, and others. Chemical composition of different organisms is given in Table 30.

Table 30. Chemical Composition of Different Organisms (% by dry weight, data of A.E.Vozbutskaya)

Organisms	Ash	Proteinous Substances	Carbohydrates		Lignin	Lipids, tannius, etc.
			cellulose	Hemicellulose plus other carbohydrates		
Bacteria	2–10	40–70	–	–	–	1–40
Algae	20–30	10–15	5–10	50-60	–	1–3
Lichens	2–6	3–5	5–10	60–80	8–10	1–3
Mosses	3–10	5–10	15–25	30–60	–	5–10
Ferns	6–7	4–5	20–30	20–30	20–30	2–10
Conifers						
wood	0.1–1	0.5–1	45–50	15–25	25–30	2–12
needles	2–5	3–8	15–20	15–20	20–30	15-20
Perennial grasses:						
cereals	5–10	5–12	25–40	25–35	15–20	2–10
legumes	5–10	10–20	25–30	15–25	15–20	2–10

Certain forms of animal life, especially earthworms, centipedes, and ants play an important role in the translocation of plant residues. There are research data to indicate the relative resistance of various organic groups to decomposition. Readily decomposed are sugars, starches, and simple proteins. Very slowly decomposed are cellulose, lignins, fats, tannins, and waxes. An intermediate position is occupied by crude proteins and hemicelluloses. All these components are usually decomposing simultaneously when fresh plant tissue is added to the soil. We may express the complete oxidation of most of the organic compounds as follows:

$$-[C,\quad 4H]+2O_2\,\frac{enzymatic}{oxidation}\,CO_2+2H_2O+Energy$$

Many intermediate steps are involved in this overall reaction. Thousands of components may be identified in SOM, the mean retention time (mrt) of which varies from a single day to hundreds and thousands of years. Living biomass consisting of roots, microorganisms, and soil fauna varies in its content from 2 to 15% of the total SOM content. Various groups and fractions of SOM are shown in Figure 75.

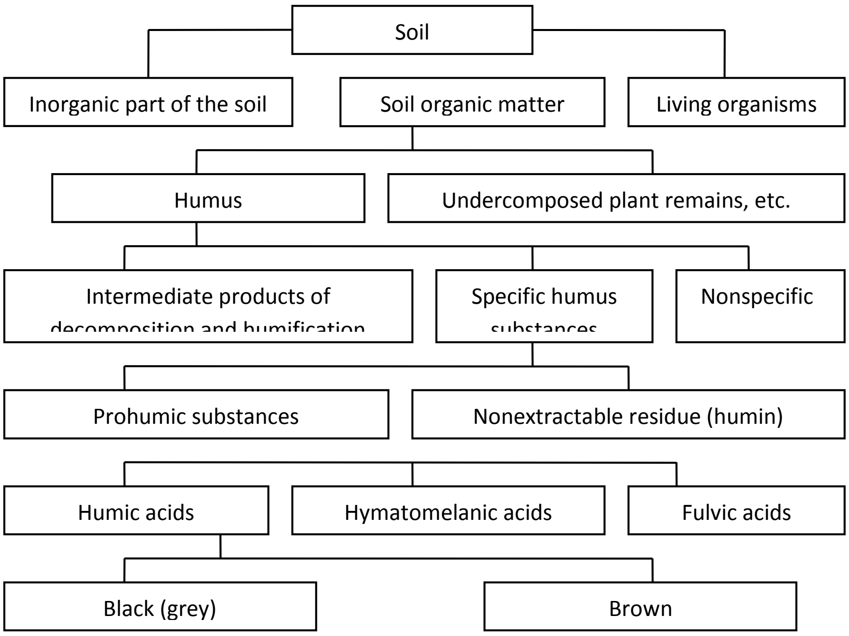


Fig 75. System of Soil Organic Matter according to D.S. Orlov, (1985)

Under decomposed plant and animal remains visible to a non-aided eye or under a magnifying glass make up 5-10% of the total nonliving organic material content in the majority of mineral soils. Humus is a complex and rather microbially resistant mixture of brown to almost black, amorphous and colloidal substances modified from the original tissues or resynthesized by soil organisms. Soil humus is divided into two large groups of substances:

1) **nonspecific organic compounds** which can be extracted from the soil and identified (sugars, aminoacids, proteins, organic bases, tannins, carbonic acids, etc). In the majority of mineral soils their content remains within several percent of the total SOM content.

2) **specific humic substances** which make up approximately 80-90% of the total SOM content in mineral soils.

A textbook of Soil Science, edited by I.S. Kaurichev (1989), gives the following definition of specific humus substances: “They are a mixture of differing in their composition and properties polymeric nitrogen-containing organic substances, and united by their common origin, some specific properties and their common – structure”. A specific properties of humus substances include:

1) specific color varying from dark brown or nearly black to reddish brown and orange for the various groups and fractions of these substances;

2) acidic character caused by the dominance of carboxyl functional groups in their molecules;

3) carbon content within 36-62%, and nitrogen content within 2.5-5.0% in the various groups and fractions;

4) the presence in molecular structures of aromatic cycles and heterocycles containing 3-6% of heterocyclic nitrogen; 5) the content of hydrolysable nitrogen within 25-35% of its total amount, and 6) the great diversity of substances by their molecular weights ranging from 700-800 to hundreds of thousands.

The classical procedure for fractionation of extracted (traditionally by 0.1 to 0.5M $NaOH$) organic matter involves acid precipitation of some fractions from the strong base extract, and subsequent dissolution of part of the precipitated material with alcohol (Fig. 76).

Fulvic acids are the most soluble group of humic substances endowed with great mobility and considerably lower molecular weights compared with the weighted average molecular weight of humic substances in general. They contain less carbon and more oxygen compared with humic acids, as well as are more light-colored than the other humic substances. They dominate in podzolic soils, rendzienes (oxisols), grey forest soils, greyzems of arid subtropic, and some tropical soils.

Humic acids are insoluble in acid group of humic substances with greater molecular weights, richer in carbon (up to 62%), and poorer in oxygen compared with fulvic acids. They are also less acidic in character. They dominate in chernozems (mollisols), chestnut soils (kashtanozems), occasionally in grey and dark grey well cultured soils.

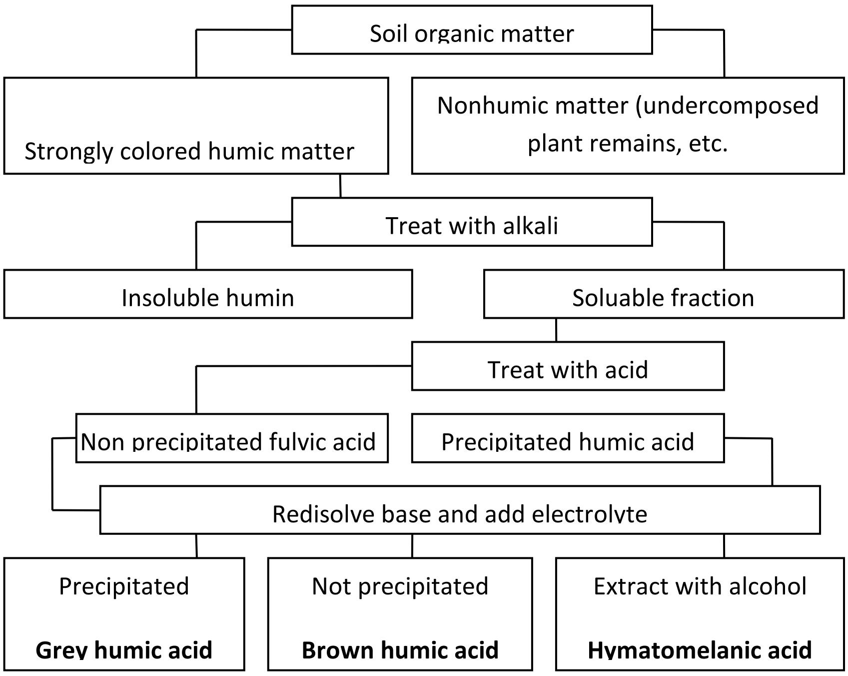


Fig 76. Fractionation of SOM (H.L.Bohn, et al., Soil Chemistry, 2nd Ed. 1985)

Humin is an insoluble in alkali (nonextractable) part of soil humus. It is evidently represented by two types of compounds: humic substances strongly bound by clays and partly decomposed organic residues without discernible signs of their former anatomical structure, rich in stable components, mainly lignins (detritic humin).

Hymatomelanic acids occupy an intermediate position between humic and fulvic acids.

14.2. Structure and Properties of Soil Humus

The degradation products of nonhumus materials undergo enzymatic and chemical reactions to form new colloidal polymers called humus. The composition and structure of soil humus is complex and incompletely known. Several researchers, however, agree on a type for humic acid. M.M. Kononova (1957) stated that “humic (or rather humus) substances are not chemically individual compounds but groups of compounds having some common structural features but varying in their nature and properties”. Elementary composition of humus acids extracted from some widespread Eurasian soils is given in Table 31.

Table 31. Elementary Composition of Humus Acids (data of L.N. Alexandrova)

Soil, location of some collecting, and sampling depth, cm	% of oven-dry ash less weight			
	C	*H*	*O*	*N*
Humic Acids				
1). Soddy-podzolic loam, Leningrad region:				
virgin forest soil	56.2	4.8	34.8	1.2
cultivated soil	56.8	4.6	34.3	4.3
2). Leached chernozem, Kursk region:				
virgin soil	60.0	3.6	32.9	3.5
arable soil	60.8	3.4	32.3	3.5
3). Krasnozem (Oxisol), Anaseuli, Georgia 0–20	59.6	4.4	31.5	4.5
Fulvic Acids				
1). Soddy-podzolic loam, Leningrad region:				
virgin forest soil	48.4	5.1	43.8	2.7
cultivated soil	46.9	4.9	45.9	2.3
2). Leached chernozem, Kursk region:				
virgin soil	45.3	4.3	47.2	3.2
arable soil	44.7	3.8	47.3	4.2
3). Krasnozem (Oxisol), Anaseuli, Georgia 0–20	49.8	3.4	44.3	2.5

A typical humus acid molecule probably consists of polymers of a basic aromatic ring structure of di - or trihydroxyl phenols linked by $-0-$, $-NH-$, $-N-$, and $-S-$ bonds, and containing both free OH groups and quinone $(O=C_6H_4=O)$ linkages. This “type” structure results in an unusually high density of reactive functional groups.

D.S. Orlov identified two structural components in any humus (humic or fulvic) acid molecule – nuclear and peripheral. Nuclear structures are stable, rich in aromatic cycles and heterocycles, less oxidized, less rich in functional groups. Peripheral structures are less “cyclic” and more “chainy”, less stable (more easily destructable), more oxidized and much richer in functional (especially carboxyl and phenol- hydroxyl) groups. The following table of data (5.4), generalized by D.S. Orlov, gives notion of the ratio of structural components in the molecules of humus acids.

Table 32. Structural and Molecular Unit Ratios in Humus Acids (%)

Structural Components	Structural Units	Humic Acids of Chernozems	Humic Acids of Podzolic Soils	Fulvic Acids of Podzolic Soils
Peripheral	Aminoacids	5.8	8.4	6.1
	Aminosugars	1.9	1.9	3.3
	Fulvic acid types of substances	6.0	7.0	–
	Total carbohydrates	29.0	27.0	44.0
	Cyclic compounds	0.5	0.4	0.2
Nuclear	Benzenecarbonic acids	25.0	10.0	6.5
	Phenolic acids	17.0	7.0	4.0
	Aminoacids	1.9	2.8	2.0
Unidentified	Unidentified	12	34	32

Humic acids are therefore richer in nuclear components whereas fulvic acids in peripheral ones. Carbon and oxygen content, acidity, and degree of polymerization – all change systematically with molecular weight. The low molecular weight fulvic acids have higher oxygen contents and lower carbon contents than the high molecular weight humic acids. The more soluble humic acids are usually responsible for the brownish-yellow color of many natural waters. Humic acids precipitate with acids and polyvalent cations, thus tending to be removed from solution.

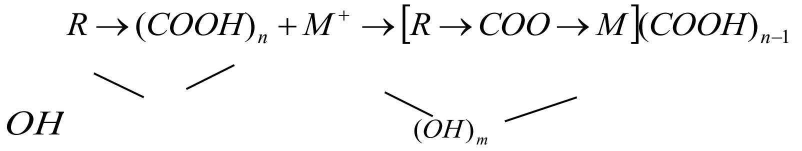
Colloidal properties. Most of the colloidal properties of SOM are due to humus. Humus is highly colloidal and is also amorphous rather than X-ray crystalline. The surface area and adsorptive capacities of humus are greater than those of the layer silicate minerals. The specific surface of well-developed humus may be as high as 900 m^2g^{-1}; its exchange capacity ranges from 1500 to 3000 mmoles (+) kg^{-1} (Bohn, et al, 1985). D.S. Orlov and other former Soviet Union students of soil humus gave values at least twice as high. The negative charge (and hence the CEC) of humus is generally agreed to be due to the dissociation of H^+ from functional groups. All charge on humus is strongly pH-dependent and its CEC increases with increasing pH. Phenolic hydroxyls begin to dissociate at pH over 8.0. CEC of fulvic acids may be twice as high as that of humic acids.

14.3. Interactions of Soil Humus with Mineral Components, Agricultural Chemicals, and Pollutants in the Soil

Humic substances, due to the multitude and diversity of their functional groups, can react with practically all the components of soils. In natural soil formation they actively interact with mineral components which attach a “desirable” stability to soil humus and cause the formation of specific soil-related accumulations of humus, macro- and microelements of plant nutrition and in a number of cases instigates soil aggregation. Other types of interactions increase the liability of mineral components favoring their leaching beyond the limits of soil profile. Humic substances actively affect the behavior of nutrient elements from fertilizers and various soil pollutants.

Sorption of humic substances by soil minerals involves a number of different mechanisms like ionic exchange, chemical sorption, complexing, intramicellar absorption of not too highly-molecular organic compounds by clay minerals with swelling crystalline structures, etc. There exists a principal possibility for the formation of hydrogen bonds and bridges through the polyvalent cations. The products formed in this way are represented by sorption complexes clay-humic compounds.

Formation of heteropolar complex salts takes place when humic and fulvic acids interact with polyvalent metallic cations like Fe^{3+}, Al^{3+}, Fe^{2+}, Cu^{2+}, Zn^{2+}, etc. The metals in such chelate complexes are no longer capable of cationic exchange. L.N. Alexandrova proposed the following scheme to describe the formation of such complex salts:

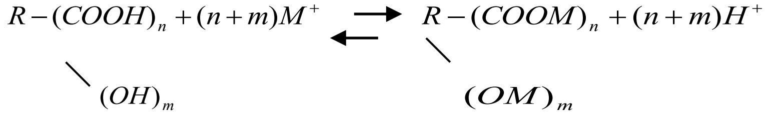


$$\underset{OH}{R \to (COOH)_n} + M^+ \to \underset{(OH)_m}{[R \to COO \to M](COOH)_{n-1}}$$

where M^+ = cations like $Fe(OH)_2{}^+$, $Fe(OH)^{2+}$, $Al(OH)_2{}^+$, etc.

It was found that in podzolic and related soils the dominant type of elements migration in the profile occurs in the form of such complexes. Some of the chelated metal elements can be absorbed by the roots. Under conditions of technogenic pollutions the polluting metals like Zn, Cu, Cd, Hg, Co, Sr, etc. can be chelated by humus components. Heteropolar complexes may interact with soluble phosphates, some pesticides, etc., changing their mobility, distribution between the phases, and availability to plants. Some soil chelates can diminish the activity at certain herbicides and bring about their degradation.

Formation of simple heteropolar salts takes place when humus acids interact with base cations like Ca^{2+}, Mg^{2+}, etc. and ammonium ions. According to L.N.Alexandrova, the following reaction occurs:



$$\underset{(OH)_m}{R-(COOH)_n} + (n+m)M^+ \rightleftarrows \underset{(OM)_m}{R-(COOM)_n} + (n+m)H^+$$

Metallic ions in this case easily dissociate and exchange with the cations of soil solution. Humates and fulvates of Na^+, K^+, NH_4, etc are soluble in water. Under natural conditions, the soluble in humates and fulvates can form, most likely, only the soils with sodic salinity. Humates and fulvates of Ca^{2+} and Mg^{2+} relatively sparingly dissolve in water and form humus accumulations in soils well saturated with bases.

There is evidence that humic substances can bind some basic pesticides like triazines. Fulvic acids can help transport some sparingly soluble pesticides.

14.4. Organic Matter in Different Soil Types

The soils of different types and different levels of cultural improvement widely differ in quantity and quality of soil organic matter. Humus content in mineral soils differs from 0.5-1% to 10-12% and more. Extensive use of naturally productive soil in crop production usually reduces their organic matter content. But low productivity forest soils may be improved in this respect by the practices of fertilizing, liming and growth of perennial grasses. Depending on the ratio between humic and fulvic acids $(C_{ha}:C_{fa})$, the following types of soil humus are identified in Ukraine: humatic ($C_{ha} : C_{fa} > 1.5$), fulvate-humatic (1.5-1), humate-fulvatic (1–0.5), and fulvatic (<0.5). Under natural conditions, which favor the accumulation of humus, humic acids dominate over fulvic ones (chernozems, sod-forming soils, meadow soils, etc). In podzolic and soddy-podzolic soils fulvic acids dominate over humic ones. Podzolized chernozems, dark-grey and chestnut soils occupy an intermediate position between these “polar” specimens. Soil reclamation and improvement in a number of cases allowed changing the type of soil humus, though more often that not, it is not easy to achieve. The organic matter content of soil is more difficult to determine then the carbon content. As a result, changes in the soil organic carbon content are used as a measure of changes in OM content. Multiplying the percentage of carbon by 1.724 is considered to equal the percentage of OM. Soil OM is constantly subject to decomposition and loss because of excessive soil tillage and shortage of organic residues in crop rotations or monoculture. Therefore, each soil and cropping situation requires the addition of a certain amount of OM each year **to maintain the status quo**. Even on nonerosive land that is converted from grassland or forest to cultivated land, a rapid loss of OM (dehumification) resulting from cropping occurs. Some quantitative and qualitative characteristics of SOM for the essential soil types are given in Table 33.

14.5. Hypotheses of Humification

Organic residues entering the soil or its surface are subject to various biochemical and physicochemical transformations. The decay of residues involves the oxidation of organic carbon by heterotrophic organisms that utilize energy from oxidation for their metabolism. The initial break up of tree trunks and large objects is carried out by mammals, termites, and earthworms. Saprophytic plants such as mushrooms obtain their energy from this partially decomposed plant matter. As the organic matter becomes more finely divided, the size of the decomposing organisms decreases.

Table 33. Some characteristics of SOM in Essential Soil Types of Eurasia

Characteristics	Essential Soil Types			
	Soddy-podzolic loams	Ordinary Chernozems	Greyzems of Arid Subtropics	Redzems (Oxisols) of Humid Subtropics
SOM content in surface horizon, %	1-4 (low and very low)	7-12 (high and very high)	1-2 (very low)	4-7 (moderate and high)
Type of humus $\left(\frac{C_{ha}}{C_{fa}}\right)$	0.4-1 (fulvatic and humate-fulvatic)	1.3-2.3 (humatic and fulvate-humatic)	0.5-1.0 humate-fulvatic)	0.4-1.0 (fulvatic and humate-fulvatic)
Total amount of SOM in 0–20 cm layer, mt/ha	30-120 (moderate, low, and very low)	200-350 (very high)	30-60 (low and very low)	120-200 (moderate and high)
C:N	12-14 (low)	13-15 (low and very low)	11-13 (low)	14-16 (very low)
Aromatic component content, %	HA FA 52-55 44-46	HA FA 57-62 42-45	HA FA 57-60 44-46	HA FA 54-57 44-47
Content of humic acids linked with Ca^{2+}, % of total humic acid (HA) content	5-7	60-70	Not determined	1-3
Nonextractable residue, % of total OM	25-35	32-40	40-50	30-45

Nonhumus organic matter includes original tissue and components and only partially decomposed materials. Nonhumus substances include carbohydrates and related compounds, proteins and their derivatives, fats, lignins, tannins, and various partially decomposed products as well as roots and plant tops. Nonhumus materials are an excellent source of food for soil microorganisms. Some plant materials (e.g. pine needles and oak leaves) are decomposed slowly and by only a few types of microbes. Shortly after fresh material comes in contact with the soil, microbes begin decomposing the residues as a source of nutrients and energy (Fig 77).

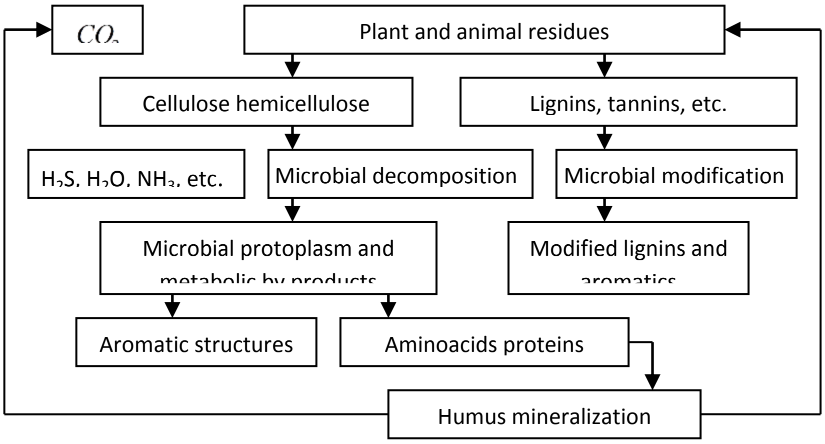


Fig 77. Organic matter decomposition and formation of humic substances. (From F.E.Bear, (Ed) Chemistry of the Soil, ACS monograph series №160. 1964, p.258).

The above figure illustrates the so-called lignino-proteinic complex hypothesis of humification. Humification is the formation of specific humus substances like humic and fulvic acids and their derivatives. Some authors (L.N. Alexandrova) consider the term humification in a wider aspect including into it not only the processes of specific polymeric humus formation but also their further transformation and degradation up to complete mineralization. The most important quantitative characteristic of humification is humification coefficient showing the ratio of humus carbon to that of initial residues from which the humus was formed. It may be expressed in decimal fraction units or in percentage. Humification coefficients are very dependent on soil hydrothermal regime, residue nature and composition (C/N), amount of residue (or organic matter) input, character of its localization in the soil determined by tillage practices, etc.

Condensation (polymerization) concept of humification was developed in various periods of time by A.G. Trusov, M.M. Kononova, and V. Fleig. Humification of plant residues is accompanied by the mineralization of some of their components to

CO_2, H_2O, NH_3 and other products. All components of plant tissues may become primary structural units of humus. The structural units undergo condensation (polymerization) with the participation of enzymes (polyphenoloxidazes) in the oxidation of polyphenols to the corresponding semiquins and quinons. The latter are capable to unite with aminoacids and other protein-related products to form a primary structure (chain) of a humus acid.

Fulvic acids, according to this hypothesis are the “progenitors” of humic acids of a higher molecular weight. Biochemical oxidation concept was put forward in the 30s of the previous century by I.V.Turin and developed in the studies of L.N. Alexandrova, according to whom humification is a complicated bio-physico-chemical process of polymeric intermediate products transportation into humus acids. Slow biochemical oxidation was thought to be of essential importance of this process. Humification is a slow and prolonged process in which linear polymeric structures become more aromatic (cyclic). Molecular weight of newly formed humic acids is higher than that of the humic acids in the soil. The system of humus substances forms as a result of oxidative acid (carboxyl) formation. This system interacts with mineral components of the soil forming a system of derivatives (simple salts, heteropolar complexes (chelates), adhesive humus-clay complexes, etc.). The system gradually disintegrates into several fractions differing in solubility, molecular weights, and the details of molecular structures. Humic and fulvic acids are formed in this way.

According to D.S. Orlov, humification can proceed by the way of condensation and by the way of slow biochemical oxidation. The first way is more typical in the formation of chernozemic and other steppe soils, whereas the second – in the formation of podzolic and related soils.

According to Selman Waksman, who emigrated to the United States in 1910, being born in the Ukrainian town of Nova Priluka, and whose contribution to soil and medical microbiology is well known to humanity, the role of microorganisms in the formation of soil humus is tremendous. He considered humification as the complexing of proteins and lignins. Selman Waksman studied chemical changes in plant materials resulting from their decomposition and the changes in C/N ratio. He investigated and described in great details the microbial decomposition of celluloses, hemicelluloses, and lignins. Waksman’s work was based on soil organic matter had pointed to a broader concept of the complex interrelationship among the many different microorganisms, especially actinomycetes and bacteria, and this led directly to studies of antibiotics.

All above mentioned views on the formation or humus substances (humification) consider both the initial row materials and final humus acids and their derivatives. A.D. Fokin in his experiments with 14C in plant residues in well developed (mature) soils with "accomplished" humus profiles showed that intermediate products of residue decomposition may be included into the ready-made humus substances as well as participate in the formation of new ones. His concept of humus formation obtained a name of "fragmentary renovation". The products of decomposition may not form new molecules of humus acids but may be involved into peripheral and, subsequently, into the nuclear structures of already existing molecules. Peripheral structures proved younger than the nuclear ones. Using the isotope indicator technique, A.D. Fokin found out that the products of decomposition were included practically simultaneously into all the groups of humic substances in ration reflecting the ratios between the structural units in the initial molecule. Thy fact explained the relative stability of the qualitative composition of soil humus in a given soil type and the difficulties encountered while trying to change this constancy by external agents.

14.6. Maintenance of Organic Matter in Cultivated Fields

As it has already been stated, soil OM is constantly subject to decomposition and loss. Soil erosion may be the reason of large OM losses from cultivated fields. Ukrainian plowland with mean OM content about 3% loses annually up to 18 mt/ha of topsoil as a result of water erosion which means the losses of SOM reaching 0.5 h mt/ha. There is evidence that in the Forest-Steppe zone of Ukraine these losses are about 1.0 mt/ha. Even on non erosive land at the Missouri Agricultural Experiment Station, as a result of 60 years of cultivation, the soil lost one third of its OM. The dynamics of soil dehumification in this case was such that the OM content decreased 25 percent during the first 30 years, 7 percent during the second 20-year period, and only 3 percent during the third 20-year period. The organic matter content of the soil was moving to a new equilibrium level.

The soil of Mykhailivska Tsilina ("tsilina" means "virgin land") is a typical chernozem (Mollisol) containing 318 mt/ha of organic carbon (548 mt/ha of OM) in the 0-100 cm layer. After 100 years of cultivation and employing the soil for the production of small grain crops, corn, sugar beet and even perennial grasses in a crop rotation the amount of organic carbon was reduced to 266 mt/ha (458 mt/ha of OM). Fig. 78 illustrates SOM Dynamics and Farm Manure Application in Ukrainian Plowland for 20 years of survey.

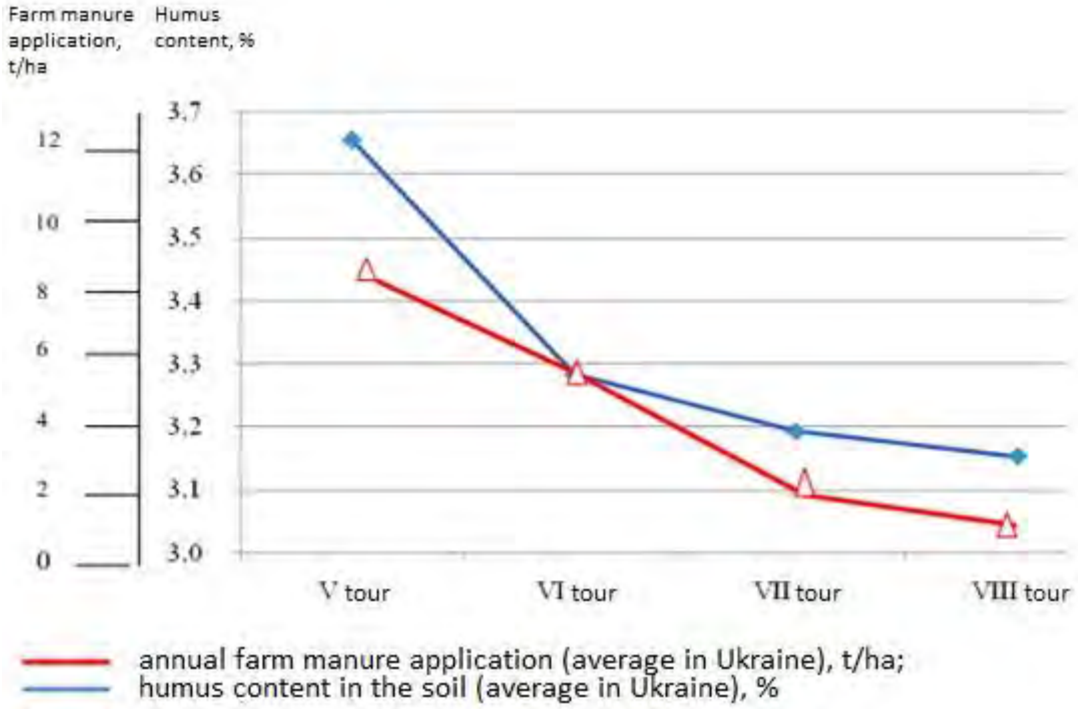


Fig. 78. SOM Dynamics and Farm Manure Application in Ukraine's Plowland (National Report on the State of Soil Fertility, 2010)

According to L.R.Petrenko (2010), the dynamics of soil dehumification, or that of the opposite process when the cultivated soils are allowed to revert to their original native vegetation, and the OM content will gradually increase toward the equilibrium level that existed prior to cultivation, can be described by the following mathematical model:

$$X_t = \left(X_0 - \frac{K_{21}\Pi}{K_1 K_3} \right) e^{-K_1 t} + \frac{K_{21}\Pi}{K_1 K_3} \tag{16}$$

where X_t is the amount of soil organic matter carbon in a given layer of soil after t years since the beginning of observation, mt/ha; X_0 is the amount of soil organic carbon in the same layer in the beginning of observation, mt/ha; K_{21} is the humification coefficient in decimal points, dimensionless; Π is an average amount of organic carbon which comes to the soil **in surface** plus root residues, organic manures, etc. annually, mt/ha; K_1 is the mineralization coefficient of soil organic matter (humus) carbon as a part of its total amount in a given layer of soil, dimensionless and expressed in decimal points; K_3 is the annual transformation coefficient for the carbon of plant residues and organic manures expressed in decimal points, dimensionless, and e - is natural logarithms basis.

To calculate the amount of organic matter (humus), X_t should be multiplied by the factor of 1.724. In a 16-year experiment in Minnesota, 5 mt of crop residues per acre (12.5 mt per hectare) annually were required to maintain the original organic carbon content of the soil at 1.8 percent. The addition of more than this amount resulted in a decline in the OM content. H.D. Foth (1990) stated that “while the land is being formed (excluding the desert land) it is virtually impossible to maintain the OM content of the virgin soil. It is therefore prudent to accept the OM content that results from economical or profitable farming”. It should not be forgotten that higher yields return more residues and maintain a higher OM content, which in turn makes still higher yields more likely. In the USA, many soil management programs focus on the production of high crop yields and proper crop residues management rather than the organic matter content of soil, per se. One of most serious causes of OM loss is soil erosion that tends to preferentially remove the organic matter fraction. High crop yields also mean more vegetative cover and reduced soil erosion.

While maintaining SOM status, it is necessary to keep in mind the significance of C/N ratio in the residues, composts, and manures. Each metric ton of straw used instead application of farm manure in Ukraine requires 10 kg of N per hectare with mineral fertilizers.

It is also important to remember that the texture of the soil, other factors being equal, seems to influence very much the percentage of humus and nitrogen present. Sandy soil usually carries less OM and N than that of a finer texture.

Poorly drained soils, because of their high moisture and relatively poor aeration, are generally much higher in OM and N than their better drained variants.

It is important to maintain soil OM above a certain (critical) level. This critical level is probably 2% for the soddy-podzolic loam-sandy soils of Ukrainian Polissya and it is probably 4.5% for chernozemic soils of the Forest-Steppe and northern Steppe in the surface layer.

As it has already been stated, the satisfactory level of OM and N is difficult to maintain in the majority of farm soils. To our mind, it is necessary to apply about 20 mt/ha of FM in the Polissya zone, 12-15 mt/ha in the Forest-Steppe zone and no less than 10 mt/ha in the Chernozemic and arid Steppe zones of Ukraine. The calculation should be made per hectare of a crop rotation. Therefore, in a 10-field crop rotation containing the fields of winter wheat and other small grain crops as well as 2 to 3 fields of sugar beet in the Forest-Steppe zone it is necessary to apply 120-150 mt/ha of FM annually (40-50 mt/ha on each of the three fields of sugar beet or corn for silage).

One of the ways to reduce OM losses is green manuring – the production of a crop for the purpose of soil improvement. These crops are easily and inexpensively established and, frequently, are grown during the fall, winter (where it is possible), and / or spring when the land would normally be unprotected by vegetation. The benefits also include erosion control. So this method of soil improvement is applied to such crops as rye, buckwheat, oats, peas, soybeans, etc.

Perennial grass crops by their nitrogen economy promote the highest possible yield of humus. Every one knows that sod begets humus.

Much can be accomplished by using a suitable crop rotation. Good soil management seeks to adjust the addition of organic residues, the physical and chemical conditions of the soil, the sequence of crops, and the losses of organic matter and nitrogen through biological activity in such a way that paying crops may be harvested without reducing the humus supply of the soil below a definite level. Any system of agriculture that fails to do this is impractical and unscientific.

14.7. Role of SOM in Soil Formation, Development of Productivity and Environmental Protection

Most obvious influences of SOM may be outlined as follows:

1) effects on soil color;
2) influence on soil physical properties;
3) encouragement of soil granulation (aggregation);
4) reduction of excessive soil plasticity and cohesion;
5) increasing soil water holding capacity.
6) ensuring high cation adsorption and the CEC;
7) accounting for 30 to 90 percent of the total absorption power of mineral soils;
8) increasing supply and improving the availability of nutrients;
9) insuring sufficient supply of N, P, and S held in organic forms;
10) extraction of nutrient elements from some minerals to make them available to plants.

But even this long list outlining the effects of SOM on soil quality is by far not complete. It ignores the environmental and sanitary functions of SOM, as well as its significance as a source of easily convertible energy.

Generally speaking, SOM's all components participate in the processes of soil formation and functioning (Fig. 79).

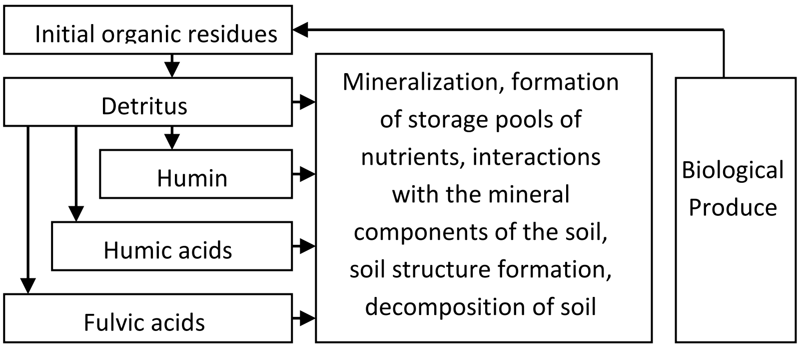


Fig 79. SOM's participation in soil formation and functioning

The functions of SOM linked with soil genesis, formation of morphological features, composition and properties are:

1) formation of a specific organic soil profile;
2) formation of soil aggregates with the participation of humus and clay-humus complexes;
3) formation of soil tilth and proper water and air-related properties;
4) formation of labile compounds capable for migration and involvement of soil mineral components into biological cycle,
5) formation of adsorptive, acid-basic, and buffering properties of soils.

Functions of SOM pertaining to the direct participation of organic compounds in plant nutrition are:

1) being a source of nutrient elements (N, P, K, Ca, S, and microelements);
2) being a source of organic (carbon) plant and microorganisms nutrition;
3) being a source of CO_2 in the soil – atmosphere layer;
4) being a source of biologically active substances, effecting plant growth and development, mobilization of nutrients and a source of vitamins and enzymes.

Functions of SOM linked with sanitation and environmental protection include the following:

1) acceleration of pesticide biodegradation;
2) retention of pollutants in soils (sorption, complexing, etc.), reduction of the input of toxicants into plant,

3) increasing migration capacity of toxicants.

Some functions of SOM are not yet sufficiently well studied or identified.

Question & Assignments

1. Name the essential sources of SOM.
2. Describe composition (system of SOM by D.S.Orlov).
3. What are specific humus substances and how they are fractionated?
4. Compare the properties of humic and fulvic acids.
5. Describe colloidal properties of soil humus.
6. What hypotheses (concept) of humification do you know?
7. How does the type of humus reflect the soil type?
8. What are the essential practices of SOM maintance in cultivated fields?
9. Describe the role of SOM in soil formation and the development of soil productivity.
10. Characterize the functions of SOM linked with sanitation and environmental protection.

Chapter 15

Plant-Soil Macro- and Micronutrient Relations

Six of the essential nutrients are used in relatively large amounts and are the macronutrients. Their content in plant tissues may usually be over 500 ppm. They include N, P, K, Ca, Mg, and S.

Micronutrients are required in very small amounts and function largely in plant enzyme system. They include B, Fe, Mn, Cu, Zn, Mo, Co, and Cl. The factors determining the amounts of micronutrients available to plants are closely related to soil conditions and plant species. A change in soil pH can change deficient situation for plants into a toxic one.

Translocation of nutrients within a plant is an ongoing process. Early in the growing season, a corn plant consists mainly of leaves, and the N exists mostly in the leaves. As the growing season continues, considerable N accumulates in the stalks and cobs. In the later parts of the growing season mobile nutrients are translocated from leaves, stalks, and cobs and are used in the development of the grain.

There is a considerable difference in the mobility of various nutrients within plants. A shortage of a mobile nutrient may cause the deficiency symptom to appear first on the older or lower leaves.

15.1. Plant – Soil Nitrogen Relations

Nitrogen plays a vitally important role in plants, being a constituent of all proteins, chlorophyll, coenzymes, and nucleic acids. Although a large quantity of N exists in the atmosphere (79 percent by volume, as inert N_2 gas), the nutrient is absorbed from the soil in the greatest quantity and is the most limiting nutrient for food production in the world.

The soil N cycle consists of N fixation, mineralization, nitrification, immobilization, and denitrification. These processes were already described in the chapter devoted to soil organisms (microorganisms). It is worth-while repeating here that N is part of nitrogenase and the N-fixing organisms also require Co.

In every ecological niche there are N-fixing organisms and systems that transfer N_2 from the atmosphere to the soil. In cultivated fields, legume-Rhizobia symbiotic fixation of N has traditionally been a primary source of N in agriculture.

At any given time, about 99 percent of soil N is organic N in OM. About 1 to 2 percent of the total organic N is decomposed and the N mineralized each year.

Many different forms of heterotrophic organisms are involved. The first inorganic or mineral form of N produced by decomposition (mineralization) is ammonia (NH_3). The process of mineralization is also called amonification.

Some NH_3 produced at or near the soil surface escapes into the atmosphere by volatilization. The same happens to some NH_3 from manure, especially when soil pH is 8 or more. NH_3 reacts with water in the soil forming NH_4^+, which is absorbed onto the cation exchange sites, becoming and exchangeable cation. NH_4^+ is about the same size as K^+, and some NH_4^+ becomes entrapped in the interlayer space of micaceous minerals (ammonium fixation). Fixed NH_4^+ has low plant availability. Some of the NH_4^+ cations are subject to nitrification:

$$NH_4^+ + 1.5O_2\,(Nitrosomonas) = NO_2 + 2H^+ + H_2O + energy.$$

$$NO_2 + 0.5O_2\,(Nitrobacter) = NO_3^+ + energy$$

Nitrification is affected by oxygen supply and soil pH. Nitrification is inhibited in overwet soils. In most well-aerated soils the dominant form of available N **is nitrate.** When soils have pH of 6 or more and are well aerated, there is an abundance of nitrifying bacteria. Nitrate is subject to leaching, whereas NH_4^+ is absorbed onto the cation exchange sites and resists leaching. N fertilizer applied in the fall is usually in the NH_4^+-form with nitrification inhibitor to reduce the possible leaching loss of NO_3^- and reduce the danger of ground water pollution.

The uptake of ammonium and nitrate nitrogen forms by plant roots and microorganisms and their conversion from inorganic to organic forms is called the **immobilization**. Soil N is subject to repeated cycling within the soil through mineralization, nitrification, and immobilization. Some plant residues materials such as wheat straw, corn stoves, and sawdust have a very small N content relative to their C content. The ratio of C to N (C/N ratio) is relatively large compared with soil humus and the leguminous plants. Bacteria require about 5 grams of C for each gram of N assimilated. Bacteria use C and N in a ratio of 5:1. When straw is added to the soil, there is insufficient N to meet the needs of the decomposing organisms. Under these conditions, decomposition of the straw is limited because of an insufficient supply of N, unless N is available from some other source. SOM may be such a source. SOM losses occur in the soil accompanied by the other kinds of soil degradation. An addition to the soil of about 10 kg of N as fertilizer per ton of straw is required to prevent a temporary N deficiency on a hectare of land area.

Anaerobic conditions favor denitrification, carried out by certain facultative and anaerobic microorganisms.

$$C_6H_{12}O_6 + 4NO_3 = 6CO_2 + 6H_2O + 2N_2$$

Nitrogen is lost from soils. But denitrification helps to prevent an excess of NO_3^- in the groundwater of irrigated fields where high rates of N fertilizer have been used. In agriculture, as long as the use of N fertilizer does not result in adding more N than crops and microorganisms can immobilize, there is little danger of groundwater pollution. But a large amount of industrially fixed N has resulted in a significant intrusion in the earth's N cycle by humans.

N-deficient plants produce low crop yields; the yellowish color in the growing season is usually an indication of N-deficiency (yellow midribs of the lower leaves of corn (maize). An excess of N causes rapid vegetative growth and dark-green leaves. The vegetation is succulent and has reduced resistance to injury from insects disease, and frost. Cereal grains develop tall, weak stalks that are easily blown over or broken during rain storms.

15.2. Plant – Soil Phosphorus Relation

The earth's crust contains about 0.1 percent of P. P, however, commonly limits plants growth. The major problem is the very low solubility of most phosphorus compounds resulting in a low concentration of phosphate ions in the soil solution at any one time. Most P occurs in the mineral apatite in igneous rocks and soil parent materials. Fluorapatite $(Ca_5(PO_4)_3F)$ is the most common apatite mineral. Apatite weathers slowly, producing the phosphate ion, $H_2PO_4^-$, which exists in the soil solution. The $H_2PO_4^-$ is immobilized when roots and microorganisms absorb it and convert to the organic compounds. This results in a significant amount of P in soils as organic P. Commonly, 20 to 30 percent of the P in plow layers of mineral soils is organic P. Organic P is mineralized by microorganisms and is again released to the soil solution as $H_2PO_4^{--}$. The phosphate ion quickly reacts with other ions in the soil solution, resulting in precipitation and adsorption to mineral colloids that convert the P to an unavailable or fixed form. Most of the phosphate ions from mineralization of organic compounds, or mineral weathering, may be converted to an unavailable form before plants have an opportunity to absorb the P and before loss by leaching can occur. Phosphate solubility is related to soil pH. The ions in the soil solution are a function of pH, a master variable of soil chemistry.

In calcareous parent materials and soils, pH is commonly in the 7.5 to 8,3 range. In these soils, apatite is the dominant P material. Apatite has very low solubility. The ions in the soil solution include Ca^{2+} from the hydrolysis of calcium carbonate. Any phosphate ions released, predominantly $H_2PO_4^-$ above pH 7.2, tend to precipitate as $CaHPO_4$ or $Ca_3(PO_4)_2$. $Ca_3(PO_4)_2$ is slowly converted to apatite. In calcareous soils most of the P tends to exist as apatite and tends to remain as apatite. P availability to plants is low.

In acid soils, there is much less Ca^{2+} and much more Al^{3+} and Fe^{3+} in solution. $AlPO_4$ and $FePO_4$ have very low solubility, resulting in low concentration of phosphate ions in solution. The formation of $AlPO_4$ ($FePO_4$) is a kind of P fixation. Phosphate ions are fixed by adsorption onto clays and other soil constituents. Fixed P dissolves slowly. The optimum pH for P availability is about 6.5, where there is the least potential for P fixation. Fresh $FePO_4$ and $AlPO_4$ are amorphous. With time, they become more crystalline and even less soluble.

In acid soils, P released by dissolution of apatite is gradually converted into $FePO_4$ and $AlPO_4$. In intensively weathered soils rich in iron and aluminum oxides, the crystalline forms of iron and aluminum phosphate become encapsulated (occluded) by iron and Al oxides.

The dominant form of P available to plants exists in the soil solution mainly as $H_2PO_4^-$ below pH 7.2 and mainly as HPO_4^{2-} above pH 7.2.

15.3. Plant – Soil Potassium Relations

The earth's crust has an average K content of 2.6 percent. Parent materials and youthful soils could easily contain 40-50 mt/ha of K. About 95 to 99 percent of this K is in the lattice of the following minerals: feldspars (microcline, orthoclase) and micas (muscovite $H_2KAl_3(SiO_4)_3$ and biotite $(H,K)_2(Mg,Fe)_2Al_2(SiO_4)_3$. Micas, especially biotite, weather faster and release their potassium much more readily than feldspars do. Weathering of micas results in the migration of K^+ out of the interlayer space along the edge of weathering mica particles. The exchangeable potassium maintains a quasi-equilibrium with the potassium entrapped or fixed. A reduced concentration of solution potassium is followed by reduced exchangeable potassium, which is then followed by the release of fixed potassium.

Plant removal and leaching tend to occur during spring and summer when the supply of available or exchangeable K is reduced. In winter, by contrast, plant uptake and leaching of K may be minimal and the release of K by weathering results in an increase in exchangeable potassium.

K fixation is affected by soil during and welting. K does not complex with organic compounds. Potassium availability is only minimally related to the soil's OM content. The most K deficient soils are those composed mostly of OM.

Potassium enhances the synthesis and translocation of carbonhydrates, thereby encouraging cell wall thickness and stalk strength.

15.4. Calcium and Magnesium as Nutrient Elements

There are many similarities between the behaviors of Ca and Mg with those of K in soil. They are all released by weathering and occur as exchangeable cations. All three elements are absorbed by roots as cations from the soil solution. Calcium minerals include calcite, dolomite, gypsum, feldspars, apatite, and amphibole. Important magnesium minerals include dolomite, biotite, serpentine, hornblende, and olivine. In contrast to K, there is no significant fixation of Ca or Mg into unavailable forms. The cations set free by weathering are absorbed onto the cation exchange sites. Equilibrium tends to be established between the exchangeable and solution forms. Mass flow commonly moves more Ca and Mg to root surfaces than plants need. Magnesium deficiency is usually associated with acidic sandy soils. When deficiencies of Ca do occur, it is usually on intensively weathered soils (like tropical ones). Magnesium is a constituent of chlorophyll. A magnesium deficiency results in a characteristic discoloration of leaves. Sometimes, a premature defoliation of the plants results. The chlorosis of tobacco, known as sand drowns, is due to magnesium deficiency.

15.5. Sulfur's Importance as a Plant Nutrient

Sulfur exists in some soil minerals, including gypsum. Mineral weathering releases S as sulfate (SO_4^{2-}) which is absorbed by roots and microorganisms. Sulfur is released as sulfur dioxide (SO_2) into the atmosphere by the burning of fossil fuels and becomes an important constituent of the precipitation. In some locations (Donbas, Kryvyi Rig), more sulfur is added to soils (via precipitation) than plants need. S accumulates in soils as organic S in plant residues and is then mineralized to SO_4^{2-}. The net effect is the accumulation of S in soils as organic sulfur. Plants depended on the SO_2 content of the air, may absorb S through leaf stomates.

The acid precipitation contributes to the development of soil acidity. As soil pH decreases, the capacity for sulfur adsorption increases. This is one reason why acid rain has only a minimal effect on many soils. S is lost from soils as sulfate by leaching and as H_2S gas, which is produced by microbial reduction of sulfates in anaerobic soils.

Plants that have the greatest need for S include: cabbage, turnips, cauliflower, onions, radishes, and asparagus. Intermediate S users are legumes, such as alfalfa, cotton, and tobacco. S deficiency symptoms are similar to N deficiency.

15.6. Micronutrients and Toxic Elements

Iron and Manganese are weathered from minerals and appear as divalent cations in solution. In this form they are available to plants. Generally, in acid soils, sufficient Fe^{2+} and Mn^{2+} exist in the soil solution to meet plant needs. In some very acid soils, Mn, and to a lesser extent Fe, are toxic because of their high amounts in solution. Deficiencies are common in alkaline soils where oxidized forms of Fe and Mn exist as insoluble oxides and hydroxides. Deficiencies are common in arid regions where many soils are calcareous and alkaline. Shrubs and flowers on calcareous soils commonly have deficiency symptoms for iron and / or manganese. Plants particularly susceptible to iron deficiency include: roses, pin oaks, azaleas, rhododendrons, and many fruits and ornamentals. Typical Fe- deficiency symptoms or iron- chlorosis appear on leaves as a yellow interveinal tissue and dark-green-colored veins.

The absence of sufficient Mn stunts tomatoes, beans, oats, tobacco, etc. The interveinal tissue turns yellow, whereas the veins remain dark-green. Iron's role in plants is in chlorophyll synthesis and in enzymes for electron transfer. Mn controls several oxidation-reduction systems and formation of O_2 in photosynthesis.

Iron is the fourth in amount in the earth's crust, following O, Si, and Al. Yet, Fe deficiencies in plants are common because the amount of soluble iron in the soil solution is frequently too low to meet plant needs. Plants increase their ability to absorb Fe from calcareous and highly alkaline soils in two ways. First, plant roots decrease pH in the rhizosphere by the excretion of H^+ that solubilizes ferric iron. Second, for monocots (grasses) **siderophores** are excreted by the roots. Siderophores (iron bodies) are metabolites secreted by organisms that form a highly stable organic chelate with iron. The **siderophores solubilize ferric iron,** which is subsequently absorbed by the roots. Microorganisms also excrete siderophores. These mechanisms enable certain plant species to consume as much iron as they need when growing on alkaline soils. But many plants are not so efficient.

Copper and Zinc are released from mineral weathering to the soil solution as Cu^{2+} and Zn^{2+}. These cations can be adsorbed onto cation exchange sites. Little Cu^{2+} exists as exchangeable copper, however, but tends to be strongly absorbed to the inorganic fraction or complexed with OM. As a result, Cu is quite immobile in soils and the Cu concentration of soil solution tends to be very low. Strongly adsorbed and nonexchangeable Cu^{2+} equilibrates with Cu^{2+} in solution. Much of Cu in solution is complexed with soluble OM of low molecular weight. Even though there is a very low concentration of Cu in solution in sufficient quantity by root interception and mass flow. Copper is not easily leached and after deficiencies in plants tend to occur on newly developed organic soils and leached about 20 to 40 kg of Cu per hectare have been applied to deficient soils, little additional Cu is needed. Zn^{2+} occurs as an exchangeable cation. It is strongly absorbed onto several soil constituents, and is complexed by OM. Zn^{2+}, however, appears to be complexed with OM to a lesser degree than Cu. It its sharply reduces availability with increasing soil pH.

Cu is a catalyst for respiration and enzyme constituent. Zn also is a constituent of enzyme systems that regulate various metabolic activities in plants. Cu is quite immobile in plants. Cu deficiency symptoms are highly variable among plants. In cereals the deficiency shows first in the leaf tips at tillering time. **The tips become white.** The leaves appear narrow and twisted. The growth of the internodes is depressed. In fruit trees, gum pockets under the bark and twig dieback occur. Zn deficiencies occur similarly to Cu on organic soils. On leached sandy soils they occur too.

Zn is commonly deficient on alkaline and / or calcareous soils. Application of P fertilizers at high rates has also been found to reduce of soil Zn are only marginally sufficient. In tobacco plants, a Zn deficiency is characterized by a spotting of the lower leaves. The white-bud of corn (maize), and the bronzing of tung tree leaves are all due to Zn deficiency.

Boron occurs in soils in some minerals like tourmaline, a borosilicate. The B is released by weathering and occurs in the soil solution mostly as undissociated with absorbed boron, boric acid. H_3BO_3. The boron in solution tends to equilibrate.

B is leached from acid sandy soils and this results in its low availability. B availability is reduced with increasing pH due to strong B adsorption to mineral surfaces (fixation). B availability is maximum in soils with intermediate pH. B deficiency tends to occur in dry weather. It is likely because of reduced movement of B to roots by mass flow in water.

Plants absorb B as uncharged boric acid. It appears to be a simple diffusion of the uncharged molecules into roots. Apical growing points and new plant leaves become chlorotic on B-deficient plants. Growing points frequently die. Many physiological diseases of plants, such as the internal cork of apples, yellows of alfalfa, top rot of tobacco, and cracked stem of celery are caused by B deficiency. Boron deficiency in sugar beets causes their heart rot.

CHLORINE. Requirement for Cl i plants is very low. There is little probability that it will ever be deficient in plants growing on soils. The presence of Cl in a fertilizer, as potassium chloride, is an additional source of Cl for many agricultural soils.

MOLYBDENUM. Weathering of minerals results Mo that is absorbed to various soil constituents. The adsorbed Mo maintains an equilibrium with Mo in solution. The solution form is mainly as the molybdate ion, MoO_4^{2-}. Mo also accumulates in SOM.

In acid soils, Mo is strongly adsorbed or fixed by Fe oxides. In many cases, Mo deficiencies are corrected by liming, owing to increased Mo availability.

Mo is needed for N fixation in legumes. When Mo is deficient, legumes show symptoms similar to those of N deficiency. In cauliflower, Mo deficiency results in the cupping of leaf edges which caused by a reduced rate of cell expansion near the leaf margin, compared with that in the center of the leaf. Leaves also tend to be long and slender, giving rise to the symptoms called whiptail. Intervenial chlorosis, stunting of plants and general paleness are also exhibited.

The Mo problem arises because forage plants have a wide range of tolerance for Mo, whereas animals do not. Legumes accumulate more Mo than common grasses do. They take up more Mo in wet than in dry soils. Molybdenosis develops in grazing animals when forage has over 10 to 20 ppm Mo.

COBALT is required by microorganisms that symbiotically fix N. This is the only known need for Co by plants. Microorganisms in the stomach of ruminants incorporate cobalt into vitamin B_{12}, and this vitamin provides these animals with Co.

People eating strictly vegetable food products are likely to have a Co-deficient diet. Cattle and sheep that are not fed legumes usually need supplementary Co. The low Co soils are very sandy, poorly drained, and have humic or organic pans (Bh horizons).

SELENIUM is not required by plants but is needed in small amounts by warm-blooded animals and humans. During the last 30 years, scientists found that the high levels of Se occurred only in soils derived from certain geologic formations with a high Se content. Another important discovery was that a group of plants, selenium accumulators, had an extraordinary ability to extract Se from the soil. They contained about 50 ppm of Se, whereas less than 5 ppm is normal. A palatable legume, Astragalus bisulcatus had a Se content about 140 times more than that of adjacent plants.

Potentially toxic polluting elements. Potential poisoning of humans by arsenic, cadmium, lead, and mercury are real concerns today. Arsenic has accumulated in soils from sprays used to control insects and weeds and to defoliate crops before harvesting. Cadmium poisoning has occurred in Japan from the dumping of mine waste into rivers, where fish ingested it. Cd also appears in some sewage sludges. Using soils for the disposal of sewage is potentially dangerous. Mine spoils may have toxic levels of Cd, but natural agricultural soils do not contain harmful levels of Cd.

Lead is discharged into the air from automobile exhausts and other sources. It eventually reaches the soil. In soil, Pb is converted to forms unavailable to plants. Any Pb that is absorbed tends to remain in plant roots and is not transported to the shoots. Soils must become much polluted with Pb before significant amounts move into the tops of plants.

Mercury is discharged into the air and water from pesticides and resulting from industrial activities. Under conditions of poor aeration, inorganic mercury is converted to methyl mercury, which is very toxic. Plants do not take up Hg readily from soils. However, soils should not be used to dispose of Hg because of a highly toxic nature of methyl mercury.

Radioactive elements. Rocks and Soils naturally contain radioactive elements. Currently, there is concern about radon. Many radioactive elements have very short half-lives. Radioactive cesium (Cs) is produced by atomic bombs and has a long half-life. Radioactive cesium appears to be fixed in vermiculite minerals much like potassium which limits its availability to plants. The soil tends to slow down the movement of radioactive cesium from the soil into plants and, later, into animals.

The contamination of soils is via the atmosphere, and radioactive elements that fall on vegetation are absorbed by the leaves. The testing of nuclear weapons prior to 1962 resulted in significant contamination of the environment. A similar contamination occurred in northern Europe as a result of the Chernobyl nuclear reactor explosion in Ukraine. Besides Cs, the soils were also contaminated with Sr. One of the approaches is to remove the top 16 inches of topsoil and replace it with uncontaminated soil.

Questions & Assiquments

Which nutrients are used by plants in relatively large amounts?

1. What are plant-deficiency symptoms and how are they related to the mobility of the nutrients within the plants?
2. What processes in the soil control the availability of nitrogen to plants?
3. Describe the phosphorus cycle in the soil.
4. Characterize the effect of pH on phosphorus and zinc availability to plants.
5. Describe soil potassium cycle.
6. Describe the role of Ca and Mg in plant nutrition.
7. In what forms do plants absorb Zn, Mn, Cu, and B? Describe the physiological roles of these elements.
8. What do you know about soil pollution with radioactive elements?
9. Which soils may be deficient in Mo?

Chapter 16

Soil Colloids and Retention Capacity

Soils are complex materials, reflecting the variability of the parent rock material and organic residues from which they form. In a simplified aspect soil may be considered as a disperse three-phase system. But the soil is a heterogeneous, polyphasic, particulate, disperse, and porous system, in which the interfacial areas per unit weight can be very large. “The disperse nature of the soil and its consequent interfacial activity give rise to such phenomena as adsorption of water and chemicals, ion (especially cation) exchange, adhesion, swelling and shrinking, dispersion and flocculation, and capillarity” (D.Hillel, 1982). The size of colloidal particles of the soil is below 0.0002 mm and it contains some 4% of such particles they determine no less than 80% of its specific surface. They influence the retention capacity of all types of soils which is very important in the development of soil fertility and plant nutrition, soil physical properties, structure, aeration, reaction, buffering, and many other properties determining the potential productivity of the soil.

16.1. Soil Colloids, Their Structure, Properties and Composition

The size of colloidal particles is within 0.2-0.001 nm. Soil colloids are of three types: 1) inorganic (mineral), 2) organic, and 3) mineral-organic. Their content in the soils may differ within the range of 1-2 to 30-40% of oven dry weight.

The origin of soil colloids is connected with the origin of soils, i.e. with the processes of weathering and soil formation. Colloidal particles (micelles) form as results of aggregation of molecules and of disintegration of larger than colloidal particles. The great specific surface and absorbing capacity of soils are due mainly to the presence of colloidal matter.

Structure and properties of colloids. Colloids are two-phased disperse systems and consist of disperse phase and soil solution. The particles of disperse phase are called colloidal micelles. They are surrounded by a diffuse double layer of hydrated ions (Fig. 80).

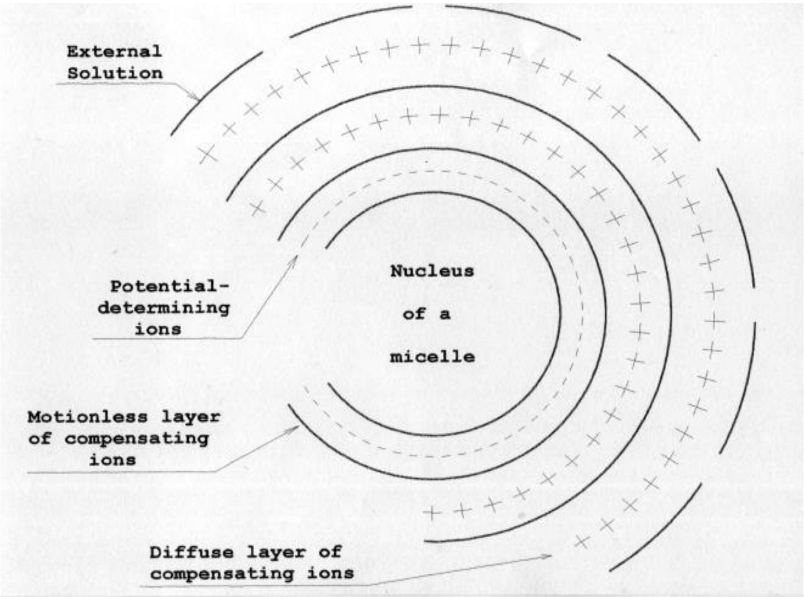


Fig 80. Typical Structure of Colloidal Micelle

If we consider the sol of H_2SiO_3 it is clear that the nucleus is composed of a certain number of electrically neutral H_2SiO_3 molecules, anions of SiO_3^{2-} form a potential-determining layer tightly held by the surface of the nucleus. So the micelle is charged negatively. $H^+(H_3O^+)$ cations from the surrounding solution function as compensating ions. A part of these cations form a tightly held by the colloidal particle compensating layer between the micelle and the surrounding (intermicellar) solution. Diffuse layer ions are not so tightly held by a colloidal particle and electrokinetic or Z-potential arises between this layer and a colloidal particle. Diffuse layer ions are considered to be the exchangeable ions which can be substituted by an equivalent number of the same charge ions from the intramicellar solution (soil solution). The electrostatic attraction of cations in H_2SiO_3 micelle is counteracted by diffusion which tends to equalize concentration differences in the aqueous phase. The cation concentration decreases with distance from the surface. So the negative charge of the colloid is neutralized by a swarm of positive ions in the solution and the diffuse double layer (DDL) is formed. The thickness of the DDL is loosely defined as the distance over which the solution concentration is affected by the colloidal charge. The solution outside the DDL is termed *the bulk solution. T*he solution near a negatively charged surface has an excess of cations and a deficit of anoions. The thickness of the DDL also decreases with an increase in charge of the cations and a decrease in charge of the anions. The total positive charge within the double layer must exactly equal the total negative charge of that region. The value of Z-potential of soil colloids changes within the range of 0 to 40-60mV. If Z-potential equals 0, the colloid is uncharged or in the state of isoelectric point. Depending on the composition of the charge-determining (potential-determining) ions, the soil colloids may be of three kinds: 1) acidoids, 2) basoids, and 3) ampholitoids. Negatively charged colloids are called *acidoids*. They contain cations in the diffuse layer (exchangeable cations). The great number of soil colloids are acidoids. Basoids are positively charged colloids. They are rare in soils. Ampholitoids can change their charge depending on the reaction (pH of the surrounding (soil) solution). Polymeric aluminum (and iron) may form colloidal micelles of (+) and (-) charge, of the general formula:

$$\left[AL(OH)_x(H_2O)^{(3-x)+}{}_{6-x}\right]_n$$

$Al(OH)_2{}^+$ ions may dominate in the soluble aluminum species at pH about 5.5. They may form a polymeric structure:

$$\left[AL(OH)_2(H_2O)_4{}^+\right]_n$$

which may serve a nucleus of a basoid with anions (like $H_2PO_4{}^-$, $HPO_4{}^{2-}$) as compensating and exchangeable ions:

$$\left[AL(OH)_2(H_2O)_4{}^+\right]_n(n-x)H_2PO_4{}^-(xH_2PO_4{}^-$$

$Al(OH)_4^-$ ions may dominate at pH over 8 and form a polymeric structure:

$$\left[AL(OH)_4(H_2O)_2{}^-\right]_n$$

which may serve a nucleus of an acidoid with cations like $H^+(Ca^{2+}, Mg^{2+}$, etc) as compensating and exchangeable ions:

$$\left[AL(OH)_4(H_2O)_2{}^-\right]_n(n-x)H^+xH^+$$

Absorption of multicharged anions can also decrease the net positive charge on hydroxy aluminum or hydroxy iron polymers, and thus increase the net positive charge of the soil-polymer mixture. The anion adsorption capacity of soils decreases with increasing pH and equals virtually zero for all anions except phosphate (and arsenate) at pH values greater than 5.5 or 6.

Soil colloids may be hydrophilic and hydrophobic depending on their relation to water. Silicate clay particles are hydrophilic. They readily adsorb water and cause the soil to swell on wetting. Parts of the water molecules are carried by the adsorbed cations, since most of them are definitely hydrated. But most silicate clays hold numerous water molecules packed between the plates that make up a clay micelle. Hydroscopic colloids practically do not hydrate. These are the nuclear part of humus molecules and polymeric hydroxides of aluminum and iron. Of silicate clays, kaolinite colloidal particles are more hydrophobic than hydrophilic. Some colloidal micelles are of crystalline structure and some are of amorphous one. Their electric charge gives to the colloids their electrokinetic properties, most essential of which are coagulation (flocculation) and peptization (deflocculation) dispersion). Colloids may exist in two states: 1) gel (colloidal precipitate) and 2) sol (colloidal solution).

Dispersion in soils occurs with monovalent and highly hydrated cations (e.g., sodium). Conversely, flocculation occurs at high solute concentrations and/or in the presence of divalent and trivalent cations (e.g., Ca^{2+}, Al^{3+}) when the double layer is compressed so that its repulsive effect is lessened and any two micelles can approach each other more closely. Thus, the short-range attractive forces (van der Waals forces) can come into play and join the individual micelles into flocs. When dispersed clay is dehydrated, it forms a dense and hard mass, or crust. On the other hand, when flocculated clay is dehydrated, it forms crumbly and loose assemblage of small aggregates. Thus, a desirable condition of a clayey soil is the flocculated one. A very good example of flocculation is provided by treating a colloidal clay suspension with a small amount of $Ca(OH)_2$. The tiny clay particles almost immediately coalesce into flocs (floccules). Because of their combined weight they sink to the bottom of the containing vessel, leaving the supernatant liquid clear.

The coagulating capacities of several cations vary with the type of colloid under consideration, but they may be ranked more or less in order of:

$$Al^{3+}\rangle Ca^{2+} and H^{+}\rangle Mg^{2+}\rangle K^{+}\rangle Na^{+}$$

Plasticity and cohesion are among the other important properties of soil colloids. Plasticity is pliability and the capacity of being molded. This property is probably due to the plate like nature of the clay particles and the lubricating and yet binding influence of the adsorbed water. In a practical way, plasticity is extremely important because it promotes such a ready change in soil structure. Cohesion is the tendency of colloidal (clay) particles to stick together. Montmorillonite exhibits cohesion to a much more noticeable degree than kaolinite or hydrous oxides do. Humus, by contrast, tends to reduce the attraction of individual clay particles to each other.

Adsorption property of soil colloids is their ability to adsorb and retain cations, anions, and the whole molecules from soil solution – may be of two kinds: 1) ionic and 2) molecular. Ionic adsorption is of an exchangeable character and is represented by cation and anion exchange reactions between the diffuse layer of soil colloids and the intramicellular (soil) solution. Molecular sorption is the retention of molecules by the components of the soil (physical retention capacity, according to K.K. Gedroiz).

Composition of soil colloids. **Mineral colloids** are represented by silicate clay minerals, colloidal forms of silica, and nonsilicate clays (Al and Fe hydroxides). Silicate clay mineral colloidal particles are laminary in shape that is made up of layers of plates, or flakes. Some of the particles are mica-like, and definitely hexagonal. Others are irregularly plate or flakelike. With some particles the edges seem to be clean cut; with other the appearance is indistinctly frayed or fluffy. In all cases the horizontal extension of the individual particles greatly exceeds their vertical dimension.

Clay particles may expose a large amount of external surface. In some clays there are internal surfaces as well. They occur between the plate-like crystal units. The external surface area of one gram of colloidal clay is at least 1.000 times that of 1 gram of coarse sand. The micelles carry a negative charge. As a consequence, thousands of positively charged ions or cations are attracted to each colloidal crystal.

Kaolins have little or no isomorphic substitution. Most of the 10 to 100 mmoles (+) kg^{-1} CEC of kaolinite has been attributed to dissociation of OH^- groups on clay edges. Halloysite is a form of kaolinite in which water is held between structural units in the basal plane, yielding c-spacing of 10 nm when fully hydrated. Most kaolin structural units, however, are held together in the basal plane by hydrogen-bonding between oxygen ions of the tetrahedral sheet and hydroxyl ions of the octahedral sheet. Surface area is limited to external surfaces and is relatively small (10 to 20×10^3 m^2 kg^{-1} = 10-20 m^2g^{-1}). Hence low colloidal activity and low plasticity of kaolins.

Smectites have relatively low layer charge. They are freely expanding. Their CEC ranges from 800 to 1200 mmoles (+) kg^{-1} (80-120 meg/100 g). The CEC is only slightly pH-dependent. Both internal and external surfaces are exposed. A total surface area is from 600 to 800 m^2*g^{-1}, with as much as 80% of the total due to internal surfaces. Montmorolonite has high colloidal activity, plasticity, and cohesion, and high swelling and shrinkage, Sources of the negative charge in silicate clays are exposed crystal edges and isomorphic substitution. At the broken edges of the silica and alumina sheets unsatisfied valences appear. The flat external surfaces of minerals have some exposed oxygen and hydroxyl groups which act as exchange sites. At high pH, the hydrogen of these OH radicals dissociates slightly and the colloidal surface is left with a negative charge carried by the oxygen. The loosely held hydrogen is readily exchangeable.

Of the so-called oxidic clays, an aluminum hydrous oxide $(Al(OH)_3)$, called gibbsite is represented by platy and hexagonal colloidal particles. It is common in intensively weathered soils. Such soils may contain abundant iron oxides, such as hematite Fe_2O_3 and goethite $(FeOOH)$. The oxide clays are abundant in many red colored soils in the humid subtropics and tropics. These soils are typically acid and in an acidic environment, some of the exposed hydroxyls on the surfaces of oxidic clays adsorb a H^+ (protonate) and form sites of positive charge ($Al - OH + H^+ = Al - OH_2{}^+$). H.D. Foth (1990) explains how a negative charge of kaolinite clay (also common in these soils) promotes the attraction of the kaolinite and oxidic clays and the formation of extremely stable aggregates or peds. Such soils have high water-infiltration rates and resist erosion. The soil can be plowed shortly after a rain. A small amount of available water is retained at field capacity. But the water is held at low tension and is easily and rapidly absorbed by roots.

Oxidic clay particles have hydroxyls exposed on their surfaces and almost no isomorphous substitution. The charge of these ampholitoides may be negative or positive, depending on pH. Increasing hydroxyl concentration, or pH, produces negative charge by deprotonating the edge hydroxyls. Under quite acid conditions with abundant H^+, the edge $Al - OH$ react with H^+ (protonate) to produce positive charge:

$$[Clay \quad edge]Al - OH + H^+ = [clay \quad edge]AlOH_2{}^+$$

Positive charge in soil adsorbs anions, thus giving the soil an **anion exchange capacity (AEC)**. There is pH where the negative and positive charges are equal; the net charge of the particles is zero. This is the **zero point of charge.** Oxidic clays have no permanent CEC, and have significant AEC in acid soils, where they are abundant.

Organic colloids are typical acidoids. A schematic formula of their micelle may be shown as:

$$\left| nR_{hum} \begin{matrix} (COOH)COO^- \\ (OH)\ldots^- \, 2H^+ \end{matrix} \right|$$

They disperse at the action of alkaline solutions and flocculate under the influence of two and threevalent cations. In soils they exist mainly as gels. They have a very high CEC. Their charge is highly pH – dependent. In acid soils the hydrogen is apparently tightly held and not subject to ready replacement by other cations. The size of particles may be at least as small as montmorillonite. Humus is not as stable as clay: it is formed and destroyed much more rapidly than clay. The main source of negative charge are carboxylic $(-COOH)$ and phenolic $(-OH)$ groups, the hydrogen of which is partially replaced by other cations. As the pH rises, first the hydrogen from the carboxyl groups and finally that from the phenolic groups ionizes and is replaced by Ca^{2+}, Mg^{2+} and other cations.

Organic-mineral colloids are acidoids and are characterized by relatively high CEC. They are found in the surface soil horizons. They are complexes of variable composition of silicate and non-silicate clay minerals with humus. Smectites and hydrous micas are very common constituents of such colloids. The process of adhesion is very important in their formation.

The colloids in any soil exist mostly as gels, hydrates to various degrees. Some colloids may exist in a free state. Others films on the surface of coarser fraction. The colloids of the first category is easily dispersed in alkaline solution and saturation of their diffuse layer with sodium. The colloids of the second category are more stable and it is not easy to disperse them.

16.2. Soil Adsorbing and Retention Capacity

Soil adsorbing and retention capacity, according to K.K. Gedroiz, is of 5 kinds: 1) mechanical retention; 2) physical adsorption; 3) physico-chemical or exchangeable adsorption; 4) chemical retention; 5) biological retention.

Mechanical retention is the ability of the soil as a porous body to retain solid particles “larger than the system of soil pores”.

Physical adsorption is, in accordance with the modern views, a molecular adsorption of the components of gases and liquids by the solid phase of the soil endowed with considerable specific surface.

Physico-chemical or exchangeable adsorption pertains to the retention by the soil of cations and anions from the surrounding soil solution by exchanging them for equivalent amounts of corresponding ions linked with soil adsorbing complex (SAC).

Chemical retention is the ability of the components of soil solution, to loam precipitates by interaction.

Biological retention is the ability of soil microorganisms and plant roots to absorb various substances from soil solution.

Soil adsorbing complex (SAC) is the pool of finely dispersed (mostly colloidal) soil particles capable for the reactions of exchangeable adsorption. The clay fraction of the soil and especially the colloidal matter is the essential sorbent.

16.2.1. Cation Retention

The adsorption of cations by the soil is performed through exchange ionic adsorption, nonexchange fixation, chemical (precipitate-forming), and biological retention. The predominantly negative charge of soil colloids retains cations in a water film on colloidal surfaces. This retention reduces the loss of Ca^{2+}, Mg^{2+}, K^+, ${NH_4}^+$ and Na^+ by leaching while keeping these cations available for plant uptake.

Exchangeable adsorption is the ability of the diffuse layer cations of the soil colloids to be replaced by an equivalent amount of the surrounding solution. Cation exchange proceeds according to the equation:

$$[SAC^{5-}]Ca^{2+}, Mg^{2+}, H^{+} + 5KCl \rightarrow [SAC^{5}]5K^{+} + CaCl_2 + MgCl_2 + HCl$$

The first studies of cation exchange were conducted by G. Thompson and J. Th. Way in Rothamsted, England, in the mid-19th century. In 1850 they, found out that cation exchange was reversible, stoichiometric, and rapid. Their results still remain valid. The order of strength of adsorption for the exchange cations is opposite to the ion replaceability, or ease of removal from specific colloids. The lyothropic series in order of relative ion replaceability could be written as:

$$Li^{+} \approx Na^{+} \rangle K^{+} \approx {NH_4}^{+} \rangle Rb^{+} \rangle Cs^{+} \approx Mg^{2+} \rangle Ca^{2+} \rangle Sr^{2+} \approx Ba^{2+} \rangle La^{3+} \approx "H^{+}" (Al^{3+}) Th^{4+}$$

The reverse order characterizes each ion’s relative affinity for or strength of retention by, montmorillonite. The strength of ionic retention and ability to replace other cations directly depends on the charge and the hydrated radius of a cation.

The distribution of major exchangeable cations in productive agricultural soils is generally $Ca^{2+} \rangle Mg^{2+} \rangle K^{+} = {NH_4}^{+} = Na^{+}$

Cation exchange capacity (CEC) is defined by H.D. Foth (1990) as the sum of exchangeable cations that a soil, a soil constituent, or other material can absorb at a specific pH; commonly expressed as milliequivalents per 100 grams or centimoles per kilogram.

It is often necessary to infer the exchangeable cation composition in equilibrium with a given irrigation water, rain water, or soil solution. Cation exchange equations are often required for such deductions.

The equation written by E.N. Gapon in 1933 is widely used (for the purpose) nowadays. It has the following form:

$$\frac{[NaX]}{\left[\frac{1}{2}CaX\right]} = K_G \frac{[Na^+]}{[Ca^{2+}]^{\frac{1}{2}}} \qquad (17)$$

where NaX and CaX is the content of exchangeable cations (Na^+ and Ca^{2+}) linked with SAC, in mg/100 g or other units.

$[Na^+]$ and $[Ca^{2+}]$ is the content of respective cations in the surrounding solution after reaching the equilibrium between this solution and the solid phase of the soil; K_G is the Gapon exchange coefficient, within 0.010-0.015 (L mmole^{-1})^{½}.

The Gapon equation is the simplest of the ion-exchange equations. It adequately predicts cation-exchange behavior over practical ranges for many soil systems.

Cation exchange is by far the most common and most important of soil reaction. Gapon was a Soviet scientist but his equation somehow did not find a sufficiently wide use in his native land. Former Soviet and Ukrainian literature often refers to Nickolsky isotherm:

$$\frac{X_i^{\frac{1}{zi}}}{X_j^{\frac{1}{zj}}} = K \frac{[i]^{\frac{1}{zi}}}{[j]^{\frac{1}{zj}}} \qquad (18)$$

The above equation is given for a pair of cations i and j; X_i and X_j are the contents of these cations in SAC and [i] and [j] – their contents (concentrations) in the surrounding solution, K, the cation exchange constant, unlike K_G, varies within a wide range of values; Z is the charge of a respective cation.

The above equation belongs to what is known in the world as the Kerr-type exchange equations. Kerr used ion concentrations in place of ion activities thus tacitly assuming concentrations and activities to be directly proportional. Nevertheless, the equation often holds fairly well for narrow concentration ranges.

Exchange reactions are rapid. The exchange step itself is virtually instantaneous. Often the rate-limiting step is ion diffusion to or from the colloidal surface. This is particularly true under field conditions, where ions may have to move through tortuous pores or through relatively thick, stagnant water films on soil colloid surfaces to reach an exchange site.

Because of their reversibility, cation exchange reactions can be driven in either forward or reverse direction by manipulating the relative concentration of reactants and products. Common techniques for driving the reactions toward completion include the use of high concentrations of exchanging cations, and maintaining low concentration of products cation by leaching or repeated washings (a procedure known as “decantation” in our soil lab practice).

Soil colloids of high charge density, that is, of high charge or CEC per unit of surface area, generally have the greatest preference for highly charged cations. Vermiculite normally retains more calcium than does montmorillonite from a mixed $Ca^{2+}-Na^{+}$ solution. Hence montmorillonite has a higher exchangeable sodium percentage than vermiculite at corresponding bulk-solution $Na^{+}+Ca^{2+}$ concentrations. The monovalent cations NH_4^{+} and K^{+} are often exception to this generalization, because of their unusually strong preference by mica and vermiculite. Raising soil pH can also change cation selectivity by increasing soil CEC and thus increasing the preferences for polyvalent versus monovalent ions. Hydrogen ion acquires one molecule of water in aqueous solutions forming a hydronium ion (H_3O^{+}, ionic radius 0.135 mm). Hydronium is adsorbed by soil colloids with greater readiness and is retained with greater strength than other monovalent ions do.

Nonexchange adsorption (fixation) of cations occurs gradually in the soil and a part of exchangeable cations may become fixed. Preferential retention of K^{+} and NH_4^{+} by vermiculite and by weathered mica edges belong to the so-called fixation reactions. Such fixation generally decreases with soil acidification, and increases with soil liming. This is attributed to the formation of aluminum hydroxide and iron hydroxide interlayers between mica and vermiculite layer lattices under acid conditions.

Lattice collapse is theoretically necessary to retain fixed cations against exchange by various extracting solutions. Fixation normally is accentuated by drying. Soils with higher OM content are more prone to fix K^{+} and NH_4^{+} than those with lower OM content. More heavy-textured soils are better cation-fixers than the less heavy-textured ones.

Exchange cation composition and the cation exchange capacity. All the soils contain Ca^{2+} and Mg^{2+} as exchangeable cations and in smaller quantities K^+ and NH_4^+. Some soils contain relatively large amounts of H^+ and Al^{3+} or Na^+. K.K. Gedroiz divided the soils into two groups: 1) saturated and 2) unsaturated with bases. Unsaturated with bases soils contain H^+ and Al^{3+} alongside with Ca^{2+}, Mg^{2+}, NH_4^+, K^+, etc. Cation exchange capacity can be determined relatively easily. In a commonly used method, the original absorbed cations are replaced by Ba^{2+}, K^+, NH_4^+ ions. Then the amount of absorbed Ba^{2+}, K^+, or NH_4^+ is determined. The CEC in most soils increases with pH. So in most cases the CEC is determined at pH of 7 or more. In Ukraine the CEC is still expressed in milliequivalents per 100 grams of oven dry soil. The value of the CEC is very dependent on soil texture and OM content. Sand and sandy loams are low in colloidal clay and are likely to be deficient in humus also. Their CEC may be as low as 2-5 meq/100 g. Finer textured soils always carry more clay and generally more organic matter as well. Their CEC are usually much higher, reaching 40-50 meq/100 g. Table 34 shows examples of the ranges of exchangeable cations to be found in soils.

Table 34. CEC Values and Major Exchangeable Cations of Selected Soils (meq/100 g)

Soil, author	Horizon	Depth, cm	Humus, %	Exchangeable cations					CEC
				Ca^{2+}	Mg^{2+}	H^+	Al^{3+}	Na^+	
Soddy-podzolic, plowland (V.K. Pestryakov)	A_{plow}	0-10	2.9	5.3	0.8	absent	0.3	absent	6.4
	A_2	29-35	0.7	3.9	0.8	» – »	0.6	» – »	5.3
	B	50-60	0.3	9.0	2.2	» – »	0.4	» – »	11.6
	C	90-100	0.2	12.5	1.6	» – »	absent	» – »	13.7
Average of agricultural soils (Netherlands)	A	0-10	–	30.2	4.98	–	0.77	2.30	38.3
Average of Calcifornian soils (H.D. Foth)	A	0-10	–	13.3	5.30	–	–	0.53	20.3

Typical Chernozem (P.G. Aderikhin)	A	0-10	9.6	46.0	9.1	–	–	–	55.1
	A	20-30	7.6	44.4	7.5	–	–	–	51.9
	AB	60-70	4.1	36.7	7.0	–	–	–	43.7
	B	80-90	3.1	36.4	7.0	–	–	–	43.4
Solonetz (alkali soil) (N.I. Usov)	A	2-8	4.6	12.8	4.4	–	–	2.9	20.1
	B_1	10-16	2.8	14.1	11.2	–	–	10.4	35.7
	B_2	24-80	0.2	20.2	6.5	–	–	7.6	34.3
	C	100-106	–	28.4	7.1	–	–	6.4	34.9
Krasnozem (Oxisol) Georgia. Tea plantation (A.I. Romashkevich)	A	0-10	6.9	2.4	1.0	0.1	9.7	–	13.2
	B_1	25-35	3.3	1.3	1.0	0.1	11.3	–	13.7
	B_2	65-85	1.7	1.3	0.7	0.1	16.4	–	18.5
	C	140-160	0.3	1.3	0.3	0.1	23.1	–	24.8

Soil properties are greatly affected by the composition of exchangeable cations. K.K. Gedroiz was one of the first researchers of soils to show the influence of exchangeable cations on soil reaction and the type of colloidal coagulation, soil physical properties and the formation of soil structure. Soils well saturated with Ca^{2+} and Mg^{2+} have a reaction close to neutral and their colloidal matter remains in the state of gels which do not yield to dispersion easily at the excess of moisture. Such soils are properly granulated and are characterized by favorable physical properties.

They are represented by chernozems, meadow and chestnut soils. The soils with an appreciable exchangeable sodium percent (over 10-15% of CEC) have an alkaline reaction unfavorably affecting soil colloids causing their dispersion. The pH of sodic soil is usually in the range of 8.5 to 10. Sodium dissociates from the colloids and small amounts of Na_2CO_3 form. Deflucculation or dispersion of colloids results in a massive or puddle soil with low water infiltration. The soil is difficult to till and the soil crust may inhibit seedling emergence. Dispersed humus coats soil particles to give them a black color. The soil is impenetrable to roots due to its dryness.

The soils containing H^+ and Al^{3+} alongside with Ca^{2+} and Mg^{2+} are unsaturated with bases to varying degree. Their reaction may be acid to varying degree. It may be toxic for many crops especially those nontolerable to the excess of Al^{3+}. Podzolic soils are a typical example. Their structure is poor and easily destructible.

Chemical retention (precipitation) of cations is their transition to the solid phase from the soil solution. Cations of Ca^{2+}, Al^{3+}, Fe^{3+}, and partly of Mg^{2+} reacting with soluble sulfates, carbonates, and phosphates may form precipitates by the following reactions:

$$\left[SAC^{2-}\right]Ca^{2+} + 2\left[SAC^{2-}\right]Ca^{2+} + Na_2SO_4 \rightarrow \left[SAC^{2-}\right]2Na^{+} + CaSO_4;$$

$$NaHCO_3 \rightarrow \left[SAC^{2-}\right]2Na^{+} + Ca(HCO_3)_2;$$

$$Ca(HCO_3)_2 \rightarrow CaCO_3 \downarrow + CO_2 + H_2O;$$

$$Ca^{2+} + H_2O + CO_2 \rightarrow CaCO_3 \downarrow + 2H^{+};$$

$$Al(OH)_3 + H_3PO_4 \rightarrow AlPO_4 \downarrow + 3H_2O$$

Biological absorption of cations is of a selective character as the living organisms (plant roots and microorganisms) absorb nutrient cations first of all. Nutrient cations include K^{+}, NH_4^{+}, Ca^{2+}, Mg^{2+}, Fe^{2+}, etc.

16.2.2. Anion Retention

Anion retention received relatively little attention researchers until recent decades. Many anions are important to present day agriculture. They include $Cl^{-}, HCO_3^{-}, CO_3^{-}, NO_3^{-}, SO_4^{2-}, HPO_4^{2-}, H_2PO_4^{2-}, OH^{-}$ Cl^{-}, and F^{-}. Besides, some micronutrients (H_2BO_3 and MoO_4^{2-}), and heavy metals (CrO_4^{2-}, $HAsO_4^{2-}$, etc.) exist as anions in soils, as well as pesticides, such as the dissociated phenoxyacetic acids (2, 4, 5-T and 2, 4-D). Anions are retained by the soil to varying extent depending on anion's nature, composition (pool) of soil colloids, and soil reaction. Some anions (like Cl^{-} and NO_3^{-}) are practically not absorbed without the participation of the living organisms. Partially absorbed are SO_4^{2-} and CO_3^{2-}, and well absorbed $H_2PO_4^{-}$, HPO_4^{2-}, and PO_4^{3-} (Table 35).

Anions are either attracted by positively charged sites on a surface, or repelled by negative charges. Layer silicates in the clay fraction of soils are normally negatively charged so that anions tend to be repelled from the mineral surfaces. Soils, however, contain a variety of solids, including the layer silicates, that develop both negative and positive charges (often simultaneous though in different locations). Anion repulsion (negative adsorbtion) may cause the equilibrium concentration of certain anions in the bulk solution to be greater than their concentration near the solid phase particles. One of the simplest explanations of negative adsorption in manuals is that the components of soil (or any other solution) increase the surface tension of this solution. Ions commonly exhibiting net anion repulsion include Cl^-, NO_3^-, and SO_4^{2-}. Anion repulsion is affected by the following factors: 1) anion charge and concentration, 2) species of exchangeable cations, 3) pH, 4) presence of other anions, and 5) nature and charge of the colloid surface. C. Mattson (1930) found out that anion repulsion in a sodium – montmorillonite suspension increased in the order: $Cl^- = NO_3 - \langle SO_4^{2-} \langle Fe(CN)_6^{4-}$. In soils that are dominated with calcium or other polyvalent cations, chemical reactions with the cations often change this purely electrostatic order.

Table 35. Anion Absorption by the Soils, meq/100 g (data of I.N. Antipov-Karatayev)

Soil	PO_4^{3-}	SO_4^{2-}	NO_3^-	Cl^-
Krasnozem (Oxisol)	74.0	7.8	Negative adsorption	Weak and negative adsorption
Podzolic soil (Spodosol)	41.0	4.2	The same	The same
Chernozem (Mollisol)	18.3	3.0	The same	The same

In soils with pH-dependent charge, lowering pH decreases the net negative charge of the system. Anion repulsion thus decreases with soil pH. Anions approaching positively charged sites on layer silicate or hydrons oxide minerals are attracted electrostatically in the same manner as cations are abstracted to negatively charged soil colloids.

Ampholitoides in acid soils causes the exchangeable adsorption of anions:

$$[SAC^{3+}]3OH + K_3PO_4 \rightarrow [SAC^{3+}]PO_4^{3-} + 3KOH$$

$$KOH + CO_2 \rightarrow K_2CO_3 + H_2O$$

Chemical and biological absorptions of anions are dominating in soils. Sulfates, carbonates, and phosphates are precipitated or are made less soluble by Ca^{2+}, Al^{3+} , and Fe^{3+}:

$$Ca(H_2PO_4)_2 + 2Ca^{2+} \rightarrow Ca_3(PO_4)_2 \downarrow + 4H^+;$$

$$Ca(H_2PO_4)_2 + 2Al^{3+} \rightarrow 2AlPO_4 + Ca^{2+} + 4H^+$$

Cation exchange can also cause the retention of soluble phosphates:

$$[SAC^{3+}]Ca^{2+} + Ca(H_2PO_4)_2 \rightarrow [SAC^{3+}]2H^+ 2CaHPO_4$$

Chemosorption of HPO_4^{2-} by the colloidal forms of R_2O_3 goes on in the following way:

$$Al(OH)_3 + (NH_4)_2HPO_4 \rightarrow AlPO_4(H^+, 2OH^-, NH_4^+) + NH_4OH$$

Phosphate is probably the most important example of specifically adsorbed anions. Many soils fix large quantities of phosphorus by readily converting soluble phosphorus to forms less available to plants. Phosphate ions are fixed by adsorption onto clay and other soil constituents. Fixed phosphorus dissolves or is released slowly into the soil solution. The optimum pH for phosphorus availability is about 6.5, where there is the least potential for phosphorus fixation. Anion exchange capacity (AEC) is the sum total of exchangeable anions that a soil can adsorb. It may be expressed in the same units as the CEC.

16.2.3. Molecular Sorption and Retention

A solute does not need to be initially charged to be adsorbed and retained by soils. This kind of adsorption was classified by K.K. Gedroiz as **physical adsorption**. It pertains to the ability of a dry soil to adsorb the molecules of water vapor from air. It pertains to any other kind of molecular (apolar, nonpolar) adsorption occurring in the soil environment. Molecules in the soil solution can become charged and then be adsorbed as cations or anions. This will be polar adsorption. They may also remain nonionic and adsorb as a consequence of polarity that produces localized charge within the molecule. **Molecules** that do not protonate or deprotonate to become charged species can still be adsorbed on soil by hydrogen bonding and van der Waals attraction.

Many organic molecules, although uncharged, are strongly retained by soils. The SOM also attracts organic molecules by providing a phase into which they can form a solid solution. Contaminated gas or water flowing through the soil is thereby purified. The soils adsorption capacity is continually renewed by microbial decay of the adsorbed molecules. The amount of molecular retention is limited by the number of exposed sorption sites, or by the amount of sorbing surface.

Adsorption isotherms are used to conveniently and quantitatively describe solute adsorption by solids at constant temperature and pressure. An adsorption isotherm shows the amount of adsorbate (solute) sorbed as a function of its equilibrium concentration. A variety of isotherm shapes is possible, depending upon the affinity of the adsorbent (solid) for the adsorbate (Fig. 81).

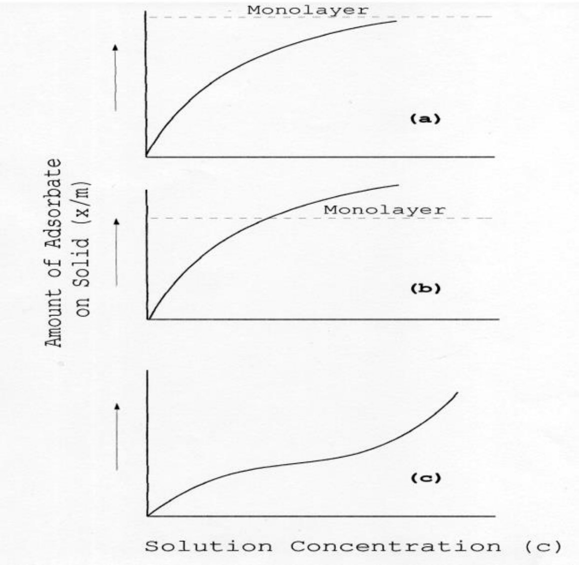


Fig 81. Typical Adsorption Isotherm Described by the a) Langmuir, b) Freundlich, and c) BET equations.

Three adsorption equations are commonly used to describe adsorption data. They are the 1) Langmuir, 2) Freundlich, and c) Brunauer-Emmet-Teller (BET) equations.

16.3. Soil Acidity and Alkalinity

Soil reaction is its very important characteristic. Studies have shown that the actual concentration of H^+ or OH^- is not very important, except under the most extreme circumstances. The associated chemical or biological environment of a certain pH is the most important factor. Some soil organisms have a rather limited tolerance to variations in pH, but other organisms can tolerate a wide pH range. Soil toxicities and nutrient availability are very much dependent on pH.

The reaction of soil solution in different soils ranges from pH of 3.5 to that of 8-9 and more. The most acid are the peat soils of the upper level swamps (pH<4). Podzolic and soddy-podzolic soils are also acid (pH 4-6). Chernozems have a close to neutral reaction and sodic soils are the most alkaline (pH may reach 11 and more). Optimum pH ranges of selected plants are given in Table 36.

Table36. Optimum pH Ranges of Selected Plants

Plants	pH	Plants	pH
Field Crops		**Fruits**	
Alfalfa	6.2-7,8	Apple	5.0-6.5
Barley	6,5-7,8	Apricot	6.0-7.0
Bean, field	6,0-7,5	Cherry. sour	6.0-7.0
Beets, sugar	6,5-8,0	Cherry, swat	6.0-7.5
Blue grass	55,-7,5	Cranberry	4.2-5.0
Clover, red	6,0-7,5	Peach	6.0-7.5
Clover, sweet	6.5-7.5	Raspberry, red	5.5-7.0
Clover. white	5.6-7.0	Strawberry	5.0-6.5
Corn (maize)	5.5-7.5	**Weeds**	
Flax	5.0-7.0	Dandelion	5.5-7.0
Oats	5.0-7.5	Foxtail	6.0-7.5
Pea, fields	6.0-7.5	Grass gnack	5.5-6.5
Peanut	5.3-6.6	Mustard, wild	6.0-8.0
Rice	5.0-6.5	Thistle	5.0-7.5
Rye	5.0-7.0	**Forest plants**	
Sorghum	5.5-7.5	Ash, white	6.0-7.5
Wheat	5.5-7.5	Aspen, American	3.8-5.5

Soybean	6.0-7.0	Beech	5.0-6.7
Tobacco	5.5-7.5	Birch, white	4.5-6.0
Vegetable Crops		Maple, sugar	6.0-7.5
Asparagus	6.0-8.0	Fir, dongles	6.0-7.0
Beets, table	6.0-7.5	Fir, balsam	5.0-6.0
Cabbage	6.0-7.5	Oak, black	6.0-7.0
Carrot	5.5-7.0	Oak, pin	5.0-6.5
Celery	5.8-7.0	Pine, red	5.0-6.0
Cucumber	5.5-7.0	Pine, white	4.5-6.0
Lettuce	6.0-7.0		
Onion	5.8-7.0		
Spinach	6.0-7.5		
Tomato	5.5-7.5		

The greater capacity of fungi, compared to bacteria, to thrive in highly acid soils was cited in the chapter on soil ecology (organisms). The pH requirements of some disease organisms are used as a management practice to control disease. Acid soil can control potato scab. Damping-off disease in nurseries is controlled by maintaining soil pH at 5.5 or less. Earthworms are inhibited by high soil acidity. Neutral reaction is characteristic of soils not containing carbonates but well saturated with exchangeable Ca^{2+} and Mg^{2+}. This reaction is most favorable for most field crops and bacteria. At soil pH about 6.2 the availability of all macro- and microelements of plant nutrition remains quite sufficient for the majority of field crops.

16.3.1 Soil Acidity

The Ukrainian standard on physicochemical terms (DSTU 3980-2000) defines soil acidity as ‘the ability of the soil to neutralize alkaline solutions and acidify water and the solutions of neutral salts". The Glossary in H.D. Foth's *Fundamentals of Soil Science* defines *acid soil* as "soil with a pH value <7.0" In Ukrainian Soil science and agricultural chemistry the so-called **active** and **potential** soil acidities are operated with.

Active acidity is defined in textbooks as "the acidity of the soil solution". Potential acidity is characteristic of the solid phase of the soil. There exist an equilibrium between the active and potential acidity expressed by the equation:

$$H^{+}\,solid\quad phase \rightleftarrows H^{+}\,soil\quad solution$$

But the dominating significance in all types of soils is attributed to the acidity of the solid phase. Active acidity of a soil solution is determined by the presence of free acids, acid salts, and the degree of their dissociation. Free mineral acids rarely occur in sizable quantities in a soil solution. In virgin swampy and podzolic soils with a high content of organic acids in soil solution their role in the creation of H^{+} increases. In the majority of soils active acidity is created by a carbonatic acid and its acid salts and is determined by the ratio:

$$H^{+} = K\frac{CO_2}{Ca(HCO_3)_2} \qquad (19)$$

As $Ca(HCO_3)_2$ is hydrolytically alkaline salt, its presence in soil solution partly neutralizes soil acidity. The presence of adsorbed H^{+} and Al^{3+} as exchangeable cations increases the acidity of the soil solution. The acidity of the soil solution is determined by titration of water extract or soil solution in milliequivalents per 100 grams of oven dry soil, or by the pH value determined in the water extracts or soil solution.

Potential soil acidity (acidity of the solid phase) has a complex nature. Its carriers the exchangeable H^{+} and Al^{3+} linked with the soil adsorbing complex. Potential acidity in Ukrainian soil science is considered as exchangeable (replaceable) and hydrolytic. Exchangeable acidity is determined by aluminum and hydrogen ions that can be removed from an acid soil by an unbuffered soil solution such as KCl. Essentially it is the sum of the exchangeable Al^{3+} and H^{+}. The Ukrainian term *exchangeable acidity* thus corresponds to the American term *replaceable acidity*. Hydrolytic acidity is higher than exchangeable one and includes it. Hydrolytic acidity is determined by aluminum and hydrogen ions that can be removed from an acid soil by the solution of a hydrolytically alkaline salt (like 1N CH_3COONa).

To understand why Al^{3+} can be the source of soil acidity it is necessary to consider the hydrolysis reactions:

$$Al(H_2O)_3^{3+} + H_2O = Al(OH)(H_2O)_5^{2+} + H_3O^+$$

$$Al(OH)(H_2O)_5^{2+} + H_2O = Al(OH)_2(H_2O)_4^{+} + H_3O^+$$

$$Al(OH)_2(H_2O)_4^{+} + H_2O = Al(OH)_3(H_2O)_3^{0} + H_3O^+$$

$$Al(OH)_3(H_2O)_3^{0} + H_2O = Al(OH)_4(H_2O)_2 + H_3O^+$$

Successive hydrolysis reactions are associated with solutions of successively higher pH, since the sink for hydronium ions (H_3O^+) increases with increasing pH. The distribution of aluminum species with pH is very important. The $Al(OH)^{2+}$ ion is of minor importance and exists over only a narrow pH range (4-5.5). The $Al(OH)_5^{2-}$ $Al(OH)_5^{2-}$ ion occurs only at pH values above those common to soils. The Al^{3+} ion is predominant below pH 4.7, $Al(OH)_2^{+}$ between pH 4.7 and 6.5, $Al(OH)_3^{0}$ between pH 6.5 and 8, and $Al(OH)_4^{-}$ above pH 8. Solid phase $Al(OH)_3$ precipitates throughout the pH range covered, whenever its solubility product is exceeded.

The very procedure of exchangeable acidity determination causes both H^+ and Al^{3+} to be the sources of acidity:

$$[SAC^-]H^+ + KCl \rightarrow [SAC^-]K^+ + HCl;$$

$$[SAC^{3-}]Al^{3+} + 3KCl \rightarrow [SAC^{3-}]3K^+ + AlCl_3;$$

$$AlCl_3 + 3H_2O \rightarrow Al(OH)_3 + 3HCl$$

It is believed by the majority of Ukrainian researchers that in soil horizons rich in humus the main role in the formation of acidity is played by H^+ whereas in more mineral horizons the role of Al^{3+} increases and may become determining in the soils formed in alitic and ferralitic cortex of weathering (oxisols and red tropical soils). Organic acids formed from the decomposition of organic residues compliment carbonic acid as a source of H^+. While reacting with the soils saturated with bases, the H^+ of these acids may replace them from the SAC:

$$[SAC^{2-}]Ca^{2+} + R(COOH)_2 \rightarrow [SAC^{2-}]2H^+ + R(COO)_2Ca;$$

$$[SAC^{2-}]Ca^{2+} + H_2CO_3 \rightarrow [SAC^{2-}]2H^+ + CaCO_3$$

Exchangeable soil acidity appears when the soil is treated with the neutral solution (1N KCl):

$$[SAC^{8-}]Ca^{2+}Mg^{2+}, H^{+}, Al^{3+} + 4KCl \rightarrow [SAC^{8-}]Ca^{2+}, Mg^{2+}, 4K^{+} + HCl + AlCl_3;$$

$$AlCl_3 + 3H_2O \rightarrow Al(OH)_3 + 3HCl$$

Exchangeable acidity may reach high values in podzolic and redzemic (oxisol) soils (pH 3-4). In soils with slightly acid or neutral reactions this form of acidity is practically absent.

Potential soil acidity reaches higher levels when the acid soil is treated with an alkaline salt like 1 N CH_3COONa:

$$[SAC^{-}]H^{2+} + CH_3COONa \rightarrow [SAC^{-}]Na^{+} + CH_3COOH$$

The quantity of CH_3COONa is determined by titration. This form of potential soil acidity (hydrolytic acidity) seems to remain important in Ukraine and other countries of the former USSR. It is expressed in milliequivalents per 100 g of oven dry soil. It is possible to determine it by the pH value of the soil – 1N CH_3COONa suspension. (40 g of air-dry soil plus 100 ml of 1N CH_3COONa and shake for 1 hour), using the formula proposed by L.R. Petrenko (2007):

$$H_h = 52.65 \times pH_{susp} \times 10^{4.73 - pH_{susp}} \qquad (20)$$

Hydrolytic acidity (H_h) is expressed here in the above stated units of meq/100 g; pH_{susp} is the pH value of the suspension obtained by the above-described procedure.

The exchangeable acidity is also expressed by pH units (pH_{KCl}), with soil to 1N KCl ratio 1:5. A soil may be very acid (pH_{KCl} < 4,5), acid (4,5-5,0), slightly acid (5,0-5,5) and close to neutral (5,5-6,0).

A soil parameter of a considerable historical importance (and in Ukraine still very important) is the percentage of base saturation, defined as

$$PBS = \frac{\sum(exchangeable\ \ Ca,\ \ Mg,\ \ Na, K) \times 100}{CEC\ \ at\ \ pH7\ \ or\, 8.2} \qquad (21)$$

pH used for base cation exchange and CEC determinations must be specified whenever this concept is to be used. In Ukraine they determine base exchange capacity (S) by the Kappen-Gilkowitz procedure and designate PBS as V, %:

$$V\% = \frac{S \times 100}{S + H_h} \tag{22}$$

Hydrolytic acidity (H_h) is determined by the Kappen procedure.

Both (pH_{KCl}) and V, % are employed to determine to what extent an acid soil needs liming:

(pH_{KCl})	V, %	Need in liming:
< 4,5	< 50	Urgent need
4.5-5.0	50-70	Moderate need
5.0-5.5	70-80	Slight need
5.5-6.0	80-90	May be a need in some crop rotations only
> 6.0	> 90	No need

Lime reacts with an acid soil by the equations:

$$CaCO_3 + H_2O + CO_2 \rightarrow Ca(HCO_3)_2;$$

$$[SAC^{2-}]2H^{2+} + Ca(HCO_3)_2 \rightleftarrows [SAC^{2-}]Ca^{2+} + 2H_2O + 2CO_2$$

In Ukraine it is believed that 1 equivalent of soil acidity requires 1 equivalent of lime $(CaCO_3)$ for its neutralization. In accordance with this assumption, liming rate (LR, mt/ha) is computed by the formula:

$$LR = 0.05 \times H_h \times h \times d_V \tag{23}$$

where h is the depth of soil layer in cm receiving lime and d_V is the bulk density of the soil in this layer, g/cm^3.

But some researcheкs do not regard the primary effect of lime to be the provision of adequate soil calcium. They attribute its main value to the provision of hydroxyl ions:

$$CaCO_3 + H_2O = Ca^{2+} + HCO_3^- + OH$$

The hydroxyl ions produced from the lime neutralize soil acidity, raise soil pH and thus provide the most important effects of the liming process. With such conception in mind LR_S will be twice as high as those computed by the above formula. In any case, increased quantities of soluble and exchangeable Ca^{2+} and Mg^{2+} are beneficial to plants having high Ca^{2+} requirements, such as legumes. Field liming reactions are generally incomplete, because of incomplete mixing and dissolution, and require considerable time. To provide maximum crop production, soil pH must be raised to near the optimum level for the crop in question. Little is gained (and much may be lost) by raising the pH to still higher levels.

Aluminum and manganese toxicities to plants are the main reason why many plants have a poor growth in acid soils. These ions are toxic to most plants and are associated with poor root penetration into acid soils (subsoils). Aluminum restricts or stops roots growth at the solution concentration as low as 1 mg L^{-1}. Plants tolerate higher levels of soluble manganese, but reducing conditions in flooded or periodically inundated acid soils can result in soluble manganese concentrations as high as 100 mg L^{-1}.

The effects of low pH on plants growth may be caused by low levels of some macronutrients and micronutrients. The availability of soil phosphate ($H_2PO_4^-$ and HPO_4^{2-}) is highly pH-dependent ($AlPO_4$ formation in very acid soils and $Ca_3(PO_4)_2$ formation in soils with high pH). The pH range of greatest phosphate availability is about 6 to 7 for most agricultural soils. Liming acid soils can either increase or decrease potassium availability. Decreased potassium availability can be attributed to increased potassium fixation in limed soils. This is similar to NH_4^+ fixation. The decreased K availability after liming can also be explained by greater quantities of K^+ being leached.

The increased soil CEC upon liming retains greater quantities of fertilizer K within the root zone and also retains it for a longer time.

Acid soils generally provide sufficient micronutrients, occasionally even toxic amounts. Molybdenum is unique among the micronutrients because it is less available to plants at low pH. Excessive liming may decrease the availability of Zn^{2+} by precipitating it in $CaZnO_2$.

16.3.2. Soil Alkalinity

H.D. Foth’s Glossary in the *Fundamentals of Soil Science* defines alkaline soil as “any soil having a pH above 7.0”. The Ukrainian Standard (DSTU 3980-2000) defines soil alkalinity as “the ability of a soil to neutralize acid solutions and make water and solutions of neutral salts more alkaline”.

Similar to acidity, there are two forms of soil alkalinity – **active** and **potential.** The above cited Standard defines active soil alkalinity as “alkalinity caused by the presence in the liquid phase (soil solution) or soil extract by water of hydrolytically alkaline salts mainly carbonates and biocarbonates of alkaline and basic metals”. Potential soil alkalinity, according to the same standard, is caused by the presence of exchangeable sodium in SAC which can be transformed into the liquid phase (soil solution) and make it more alkaline, SAC in this definition being **the soil adsorbing complex**.

Active alkalinity is determined by the soil extract (soil: distilled water = 1:5) titration with the H_2SO_4 solution (0.05N in Ukraine) and expressed in milliequivalents per 100 grams of oven-dry soil:

$$2Na_2CO_3 + H_2SO_4 \xrightarrow{phenolphtalein} 2NaHCO_3 + Na_2SO_4$$

$$2NaHCO_3 + H_2SO_4 \xrightarrow{phenolphtalein} Na_2SO_4 + 2H_2O + 2CO_2$$

Alkalinity caused by the presence of HCO_3^- in soil solution is “mild” and the pH does not exceed 8.4. Alkalinity caused by the presence of CO_3^{2-} in soil solution is high (pH > 8.4), as belonging to Na_2CO_3, CO_3^{2-} causes the following hydrolysis reaction:

$$Na_2CO_3 + 2HOH \rightleftarrows H_2CO_3 + 2Na^+ + 2OH^-$$

Sodium carbonate may appear in a soil as a result of exchange reactions involving adsorbed sodium ion, or as a consequence of action of sulfate-reducing bacteria in anaerobic conditions by the equations:

$$Na_2SO_4 + 2C = 2CO_2 + Na_2S$$

$$Na_2S + CO_2 + H_2O = Na_2CO_3 + H_2S$$

Potential alkalinity is characteristic of soils containing considerable amounts of exchangeable Na^+ in SAC (over 5-10% of the CEC). Reacting with carbonic acid such a soil is likely to produce Na_2CO_3:

$$[SAC^{2-}]2Na + H_2CO_3 \leftrightarrows [SAC^{2-}]2H^+ + Na_2CO_3$$

Soil alkalinity, like soil acidity, can be expressed by the pH values of soil solution or soil extract in water. The following ranges of soil alkalinity are identified in the

Ukrainian soil surveys: 1) slightly alkaline (pH of 7.2-7.5), 2) alkaline (7.5-8.5), and 3) very alkaline (> 8.5).

Some alkaline especially sodic soils are very unfavorable for plants growth and development. In the USA sodic soils are those whose exchangeable sodium percentage (ESP) is 15 or more. According to K.K. Gedroiz, sodic (alkali, solonetzic, etc) soils may develop when the leaching of a saline soil results in high exchangeable sodium and low exchangeable calcium and magnesium.

pH is usually in the range of 8.5 to 10 and more. Sodium dissociates from the colloids and small amounts of Na_2CO_3 form. Defloculation, or dispersion, of colloids occurs. The soil structure breaks down. The soil becomes massive or puddled with low water infiltration, difficult to till (sticky to implements), etc. According to I.N. Antipov-Karataev, the extent of soil sodicity, depending on ESP, may be as follows:

Extent of Sodicity	ESP
Nonsodic (nonsolonetzic)	< 5
Slightly sodic (slightly solonetzic)	5-10
Moderately sodic (moderately solonetzic)	10-15
Strongly sodic (strongly solonetzic)	15-20
Sodic alkali (solonetz)	> 20

Some solonetz soils in the chernozemic Steppe zone are a real picture to see: organic matter in the soil disperses along with the clay, and the humus coats soil particles to give them a black color. A characteristic soil structure with whitish rounded heads and black downs may occur.

The basic treatment of sodic soils is the replacement of exchangeable sodium with calcium of the finely ground gypsum $(CaSO_4 \cdot 2H_2O)$:

$$\left[SAC^{2-}\right]2Na + CaSO_4 \rightleftarrows \left[SAC^{2-}\right]Ca^{2+} + Na_2SO_4$$

Gypsum gradually improves the soil in many ways and increases the yields of crops. Na_2SO_4 may need leaching from the soil as it is slightly toxic to agricultural crops. The amount of gypsum needed to replace the exchangeable sodium is the gypsum rate (GR). In the Ukrainian soil amendment tradition the most universal formula for GR computation is the following:

$$GRmt/ha = 0.086(Na - 0.05CEC) \times h \times d \qquad (24)$$

where 0.086 is the milliequivalent of $CaSO_4 \cdot 2H_2O$ in grams; Na – is the exchangeable sodium content in meq/100 g; h – the depth (cm) of the layer in which gypsum is incorporated, and d is the bulk density of the soil in this layer, g/cm^3.

16.4. Soil Buffer Capacity

The buffer capacity is the ability of a soil to counteract the changes in the reaction of the soil solution. This capacity is to a great extent determined by the ability of ions associated with the solid phase to buffer changes in ion concentration in the solution phase. In acid soils, buffering refers to the ability of the exchangeable H^+, Al^{3+}, and hydroxyl-aluminum cations to maintain a certain concentration of H^+ in a solution. The amount of H^+ in the soil solution of a soil with pH of 6.0, for example, is extremely small compared to the nondissociated H^+ adsorbed and the amount of aluminum that can hydrolyze to produce H^+. Neutralization of the active or solution H^+ results in rapid replacement of H^+ from a relatively large amount of H^+ associated with a solid phase. Thus, the soil has shous great resistance to undergo a pH change.

Actually there are different mechanisms for soils to resist pH change whether acidity or alkalinity is introduced in one form or another.

Figure 82 depicts a hypothetical soil pH – buffer capacity curve.

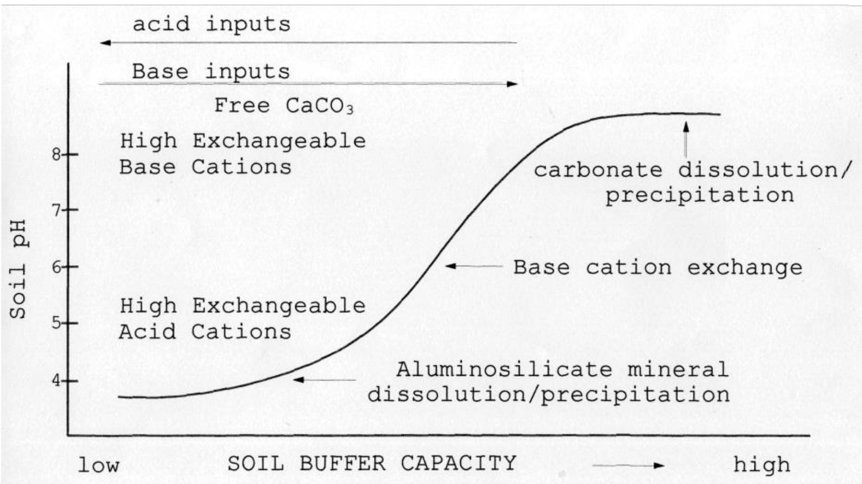


Fig. 82. General relationship between soil pH and the quantity of acid or base added. The inverse of the slope of this curve is a measure of soil buffer capacity (H.L. Bohn et al., *Soil Chemistry*, 1985).

Soils well saturated with bases have a high buffering capacity against the acidification of soil solution:

$$[SAC^{4-}]Ca^{2+}(Mg^{2+}) + 2H_2CO_3 \rightleftarrows [SAC^{4-}]4H^+ + CaCO_3 + MgCO_3$$

Soils unsaturated with bases have a high buffering capacity against the agents of soil solution alkalinity:

$$[SAC^{2-}]2H^+ + Na_2CO_3 \leftrightarrows [SAC^{2-}]2Na^+ + H_2O + CO_2$$

The composition of soil solution also affects buffering capacity of the soil owing to the presence of carbonic and weak organic acids and their salts with Ca^{2+}, Mg^{2+}, K^+, Na^+, etc. Soil buffering capacity must be taken into account while solving the problems of fertilizing and soil amendment.

Questions & Assignments

1. Speak on the structure, properties, and composition of soil colloids.
2. What is flocculation and dispersion? Coagulating capacity of cations.
3. Describe silicate clays as colloids. What is the CEC?
4. Describe colloidal properties of the SOM.
5. What is AEC and in what units it may be measured?
6. What are the kinds of soil absorption capacity identified by K.K. Gedroiz?
7. How is the CEC effected by pH of the bulk solution?
8. Analyse the Gapon equation and give reasons for its wide application.
9. Speak on the nonexchange adsorption of cations.
10. Give examples of chemical retention (precipitation) of cations and anions.
11. Prove the importance of molecular adsorption for soil productivity.
12. Speak on the forms and neutralization of soil acidity.
13. Speak on the forms and neutralization of soil alkalinity.
14. Name soil properties most strongly effecting its buffer capacity. Explain why.

Chapter 17

Soil Structure

Soil particles may exist freely not cemented with each other or be united into structural aggregates (crumbles, grains, nuts, prisms, clods, and peds) of different shape and size under the influence of various agents. Soil aggregates may unite with each other forming a distinctly structured pattern. The soil then behaves as a composite body with properties depending on the manner in which the various particles are packed and held together in a continuous spatial network which is commonly called *the soil matrix* or *fabric*.

The arrangement and organization of the particles in the soil is called *soil structure***.** The aggregates of different shape and size are also different in their composition and resistant to various stresses.

Sandy soils do not have any structure – they are granular. Loamy and clayey soils may be structured or not structured (massive) or consolidated. We have already considered the morphological and classification aspects of soil structure (Chapter 3). In Chapters the soil structure will be characterized in the aspects of its agronomical value. The practice of soil cultivation made it abundantly evident that soil structure affects soil physical and physico-mechanical properties, conditions of soil tillage, water- and air- related soil regimes, soil fertility, and plants growth.

17.1. Agronomical Importance of Soil Structure

The quality of soil structure is determined by the size of the aggregates, their shape, strength (to crushing and to raindrop impact), porosity, etc. The most agronomically favorable are the macroaggregates (10-0.25 mm), close to spherical in shape (grains and clods), endowed with considerable porosity (over 45% in volume), mechanically strong and resistant to the action of water (raindrop impact and surface flow). If the aggregates are mechanically strong (shear stress resistance) and not subject to disintegration in water, they are capable to be preserved to a sufficient extent even under the impact of numerous tillage and traffic passes through the field. Otherwise the clods will be easily disintegrated into smaller and smaller particles by the action of tillage, raindrop impact, and the surface flow of water during the rainfalls and irrigation practices. The soil will easily become structureless, subject to puddling and crust formation.

Not any “waterproof” structure is agronomically favorable. Soil aggregates must be loosely packed. Their “spatial network” should allow sufficient total size porosity (55-65% by volume). Soil water permeability should be sufficiently high (say about 100 mm per hour). Soil pores must be penetrable for plant roots and rootlets. In compacted soils the aggregates are not loosely packed, and total soil porosity becomes lower than it is necessary (30-40% by volume). The aggregates may be water-resistant but only due to weak penetration of water into them. Such structure is not agronomically favorable. Usually such morphological forms of soil aggregates as plates, leaves, prisms columns, and blocks do not belong to agronomically favorable ones. A cross section of cultured and well structured soil is given in Figure 83.

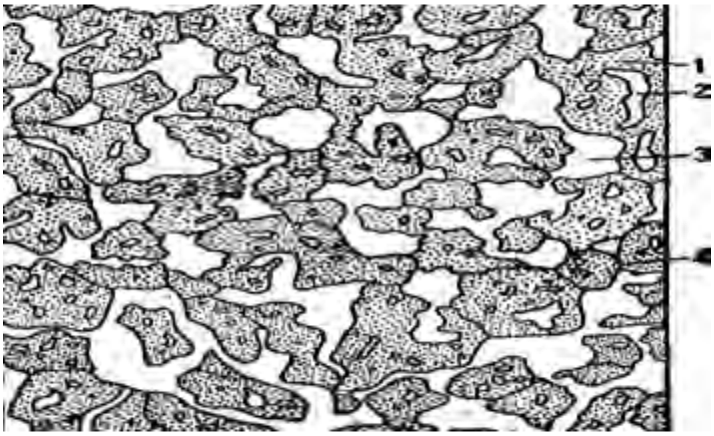


Fig 83. Porosity of a soil with agronomically favorable structure (by N.A. Kachinsky):

1) fine capillary pores;
2) moderate-sized pores in clods;
3) large pores between the aggregates (usually filled with air);
4) capillary pores between the aggregates, usually filled with water.

Aggregate stability is of great practical importance. Organic compounds of various kinds are known to have stabilizing properties. The kind of clay also influences aggregate stability. Kaolinite-dominated granules are more stable. Other inorganic compounds, such as iron oxides, have a definite cementing or binding effect. As a general rule, the larger the aggregates, the lower their stability.

Agronomical value of soil structure is, generally speaking, determined by its positive effects on the following soil properties and regimes:

1) physical properties: porosity and bulk density;
2) water, air, temperature, red-ox, microbiological, and nutritive regimes;

3) physico-mechanical properties of soils: shear resistance, resistance to tillage, penetrometer resistance, crust formation, etc.;

4) soil resistance to erosion (by water and wind).

Agronomically favorable structure creates a favorable ratio between aeration porosity and capillary (water retaining) porosity in the soil.

D. Hillel in his *Introduction to Soil Physics* (1982) states that "since soil particles differ in size, shape, and orientation, and can be variously associated and interlinked, the mass of them can form complex and irregular configurations which are in general exceedingly difficult if not impossible to characterize in exact terms". Soil structure is strongly affected by changes in climate, biological activity, and soil management practices. But as soil structure affects the retention and transmission of fluids in the soil, including infiltration, water-holding capacity, and aeration, as soil structure influences the mechanical properties of the soil, and may also affect such phenomena as germination, root growth, tillage, overland traffic, and erosion. The knowledge of soil structure relationships is essential for efficient management.

Large cohesive blocks over 10 mm in size (megastructure) and small granules of dust less than 0.25 mm in diameter are not desirable for agronomical purposes. The formation and maintenance of stable aggregates 10-0.25 mm in size is the essential feature of *soil filth*, a qualitative term used by agronomists to describe that highly desirable, yet unfortunately elusive, physical condition in which the soil is an optimally loose, friable, and porous assemblage of aggregates permitting free movement of water and air, easy cultivation and planting, and unobstructed germination and roots growth.

In soils with an appreciable content of clay, the primary particles tend under favorable circumstances, to group themselves into structural units (primary aggregates). Primary aggregates combine into secondary aggregates and so on. The visible aggregates, which are generally of the other of surface millimeters in diameter are the most agronomically favorable. But in soils with excessive wetness some larger aggregates (5-10 mm) are preferable.

A prerequisite for aggregation, according to D. Hillel, is that the clay be flocculated. However, flocculation, using a mathematics expression, is a necessary but not a sufficient condition for aggregation. In addition to it a proper cementation of particles is needed.

In soils with proper structures there is no antagonism between soil water and soil air. Even at field capacity the aeration porosity is over 15% by volume. Such a condition of soil filth favors the useful microbiological processes, whereas denitrification, reduction processes with the excess of toxic Fe^{2+}, Mn^{2+}, etc. do not take place.

Structureless (for example compacted) soil slowly absorbs water, so that much of it is lost by evaporation or surface runoff. Unbroken capillaries filled with water discharge (lose) it rather rapidly. When dried and cracked, the same rapid losses of water from the soil occur.

Soil microstructure may also be of use if its content is a good compliment for the favorable macroaggregates. But the microaggregates less than 0.01 mm in size impede air and water circulation and favor the rapid losses of water. Surface layer containing over 60% by volume of soil particles and aggregates over 1 mm in diameter is resistant to wind erosion. It is the aggregates exposed at the soil surface which are the most vulnerable to destructive forces. The surface aggregates which collapse and slake down during wetting may form a slick layer of dispersed mud, sometimes several centimeters thick, which clogs the surface micropores and thus tends to inhibit the infiltration of water into the soil and the exchange of gases between the soil and the atmosphere. Such a layer is often called *a surface seal*. Upon drying, this dispersed layer shrinks to become a dense, hard crust which impedes seedling emergence by its hardness and tears seedling roots as it cracks. Soils vary in the degree to which they are vulnerable to the externally imposed destructive forces. But aggregate stability is a measure of this vulnerability. It expresses the resistance of aggregates to breakdown.

17.2. Formation of Soil Structure

According to I.S. Kaurichev's textbook still very popular in Ukraine, two essential processes are important in the formation of soil macrostructure: 1) mechanical disintegration of soil mass into aggregates (clods) and 2) formation of strong soil granules resistant to the action of water. These processes occur under the influence of physico-mechanical, physicochemical, chemical, and biological agents of soil granulation.

Physico-mechanical (and purely physical) factors of the formation of soil structure. They cause the disintegration of soil mass under the impact of changing pressure or mechanical action. Soil mass may change its volume (and pressure) with alternating drying and wetting, freezing and thawing, root pressure, action of moles and other tunneling animals (earthworms), shoveling effects of tillage implements, etc. For many reasonable students of soil structure the mechanisms of its formation still remain exceedingly complicated and rather obscure. But many allow themselves to speculate on this score. At any rate it is useful to remember that tillage should be carried out at soil moisture as close to the lower index of soil plasticity as possible. This will prevent an excessive destruction of soil structure. But no tillage ever created any structure.

Physico-chemical agents – coagulation (flocculation) and cementing action of soil colloids really are important. The soil must be well saturated with certain exchangeable cations like Ca^{2+}, Mg^{2+} (but not excess of Mg^{2+}), Fe^{3+}, and Al^{3+}. Exchangeable Na^{+} is in the way of soil colloid flocculation being a dispersing agent breaking up soil structure. The same pertains to the excess of K^{+}, NH_4^{+}, and other alkaline cations. Organic colloids, especially calcium humates, are good cementing materials for soil aggregates. The influence of decaying organic matter from the microorganisms and other forms of life, as well as some root exudates, seem to favor soil granulation.

Some glueing and cementing action can be expected from purely chemical agents of soil granulation. Unsoluble precipitate formation ($CaCO_3$, $Fe(OH)_3$, $Al(OH)_3$, $MgSiO_3$, etc) helps to cement soil particles and microaggregates into clods and other forms of soil structure. Temporary excessive wetness may instigate iron to become a cementing agent. Reduction processes during wetting favor iron (Fe^{2+}) compounds dissolution whereas subsequent oxidation at drying causes the precipitation of Fe^{3+}-compounds which cement mechanical soil particles and the aggregates of the smaller order. Such aggregates may be common in drained swampy soils but are especially abundant in redzems (oxisols). But the aggregates cemented with iron (or aluminum) compounds, though resistant to the action of water, have low porosity (<40% by volume). $Fe(OH)_3$ and $FeOOH$ fill the pores in the aggregates.

The essential role in the formation of soil structure belongs to biological agents that are vegetation and the organisms inhabiting the soil. Perennial grasses have a well-developed root system extending in the soil in all directions. Root residues and root exudates favor the formation of soil humus well saturated with calcium.

A mere visual inspection of virgin chernozems presents the abundance of “buckwheat grains” and “small peas” of soil aggregates. Meadow and alluvial soils are always “grainy”. Earthworm activity creates “coprolites” or earthworm casts (pellets). These casts are resistant to water action but their internal porosity may be low. Colloidal products of microbial decomposition also are cementing materials for the formation of soil aggregates.

The agents of aggregate formation and development of their stability are difficult to differentiate as they are strongly interconnected. N.I. Savvinov studied the stability of macrostructure in the surface (mineral) horizons of virgin soils from the tundra to the semidesert zone. This stability was the highest for the steppe chernozems (mollisols), gradually decreasing to the north through the forest-steppe, forest, and tundra zones and more rapidly decreasing to the south through the arid steppe, semidesert, and desert zones. In humid subtropics with red-colored oxisols this stability was very high because of iron and aluminum hydroxide cementation of soil aggregates.

17.3. Destruction and Renewal of Soil Structure

Soil structure is very dynamic. Determining the momentary state of aggregation of a soil at any particular time might not suffice to portray the soil's true structural characteristics as they may vary dynamically over a period of time. The aggregates form, disintegrate, and reform periodically. The disintegration of soil structure is believed by many of Ukrainian pedologists to be caused by **mechanical, physicochemical, and biological processes.**

Repeated traffic, particularly of heavy machinery, tends to crush the aggregates remaining on the surface, and to compact the soil to some depth below the surface. Raindrops develop a high velocity near the soil surface (up to 8-10 m/sec) and hit the soil aggregates with a tremendous force, causing their gradual or rapid break-up. Soil moisture near the lower plasticity index is in Ukrainian soil science called the "physical ripeness" for tillage operations. With such soil moisture, during plowing or other tillage operations, the disintegration of aggregates is minimal.

Physicochemical causes for soil aggregate disintegration are linked with the dispersion of soil colloids brought about by the replacement of Ca^{2+} and Mg^{2+} from the SAC with Na^{+}, K^{+}, NH_4^{+} and other dispersing cations.

D. Hillel (1982) stresses the fact that the concept of aggregate stability should be applied first of all to the destructive action of water. The very wetting of aggregates (to say nothing of raindrop impact) may cause their collapse. The bonding substances dissolve or weaken and as the clay swells and possibly disperses, the aggregates disintegrate. The abundance of coarse silt fraction (0.05-0.01 mm) in the Ukrainian loess soils make the aggregates yield to disintegration in water. If wetting is nonuniform, one part of the aggregate swells more than the other. The resulting stress may fracture the aggregate. It is worth saying again that raindrops and flowing water provide energy to detach particles and transport them away. The classical and still most prevalent procedure for testing the water stability of soil aggregates is the wet sieving method proposed by A.F. Tiylin (1928) and developed by N.I. Savvinov (1928).

All the practices of soil amendment such as liming of acid soils and gypsing of sodicity affected ones are very operative in increasing the stability of their structures.

Biological causes for the disintegration of soil structure are linked with the mineralization of soil humus – the essential cementing material in aggregate formation. Reproduction of soil structure in crop production systems is achieved by owing to the practice of agrotechnologies. It is also possible to use specific chemicals for the purpose.

The cultivation of perennial grasses, soil tillage at the state of agrophysical ripeness, liming of acid soils and gypsing of sodic or sodicity affected soils, application of organic manures and mineral fertilizers in proper rates and rations – all such practices have to improve or preserve soil structure. It is certainly impossible to restore the aggregation of the virgin chernozems but it is a must in crop production systems to reduce the disintegration of soil aggregates to tolerable rates. Annual crops can also improve soil structure but to a much lesser extent than perennial grasses. The extensive cultivation of row (intertilled) crops destroys the soil structure and enhances the losses of SOM at an accelerated rate, especially if such crops are grown on the sloping land. Soil erosion then may exceed tolerable rates many times.

The traditional and still widely accepted practices of tillage are based on a series of primary cultivations followed by the secondary ones. In the process, energy is often wasted and natural soil structure may be destroyed.

The more modern approach to soil structure management conceives of a field typically planted to row crops as consisting of at least two distinctly different zones:

1) **A planting zone,** where conditions are to be optimal for sowing and conducive to rapid and complete germination and seedling establishment.

2) **A management zone** in the interrow areas, where soil structure is to be raw and open, allowing maximal intake of water and air and minimal erosion and weed infestation**.**

Recent trends in tillage research are aimed at minimizing tillage operations and travel. This approach in numerous variations underlies the methods of “wheel-track planting”, “precision tillage”, “zero tillage”, “minimum tillage”, etc. It is quite natural that some soils require little if any tillage at all to serve as favorable media for crop growth.

The use of chemical polymers of acrylic, maleinic and other acids called “crylliums” remains very expensive for most practical purposes. The rates of such chemicals should be no less than 0.001% of the amended soil weight.

Questions and Assignments

1. Give definitions to soil structure per se and agronomically favorable soil structure
2. What forms of aggregate stability do you know?
3. Name agronomic advantages of good soil structure.
4. Describe the mechanisms of soil aggregate formation.
5. Describe practices for the management of soil structure.

Chapter 18

Physical Properties of Soils

Physical properties of soils, in according to Ukrainian soil science, include soil structure, water-related soil properties, and the condition of soil air and aeration, soil temperature, thermal regime, and heat balance, general physical and physico-mechanical properties of soils. This chapter characterizes general physical and physico-mechanical properties of soils.

18.1. General Physical Characteristics of the Soil

"Soil physics is the branch of soil science dealing with the physical properties of the soil, as well as with the measurement, prediction, and control of the physical processes taking place in and through the soil" (D. Hillel, 1982). General physical characteristics of the soil pertain to it as a disperse three-phase system, consisting of the solid phase (or soil matrix), the liquid phase (soil solution), and the gaseous phase in the soil (soil air). The relative proportions of the three phases in the soil vary continuously, and depend upon such variables as weather, vegetation, and management.

General physical characteristics of soil include 1) density of soil solids (mean particle density); 2) dry bulk density (bulk density), and 3) porosity.

Particle density, ρ_S (D in Ukraine)

$$\rho_S = \frac{M_S}{V_S} \tag{25}$$

where M_S is the mass of soil solids, and V_S is the volume of soil solids. ρ_S (D) is expressed in gm/cm^3 (mt/ha).

In most mineral soils, the mean density of the particles is about 2.6-2.7 gm/cm^3, and is thus close to the density of quartz, which is often prevalent in sandy and silty soils. Aluminosilicate clay minerals have a similar density. The presence of iron oxides, and of various heavy minerals, increases the average value of D, whereas the presence of OM lowers it. Sometimes the density is expressed in terms of the specific gravity, being the ratio of the density of the material to that of water at 4 °C and atmospheric pressure.

D for mineral soils usually remains within the range of 2.40-2.80 gm/cm^3. Soddy-podzolic soils poor in OM content have it within the range of 2.65-2.70;

subtropical and tropical soils are rich in R_2O_3 (Al_2O_3 + Fe_2O_3) – 2.70-2.80. Peats may have particle density of 1.40-1.80 gm/cm^3.

Bulk density, ρ_b (d in the Ukrainian scientific tradition)

$$\rho_b = d = \frac{M_S}{V_t} \qquad (26)$$

where V_t = the total volume of the soil.

The dry bulk density expresses the ratio of the mass of dried soil to its total volume (solids and pores together). Obviously, ρ_b (d) is always smaller than ρ_S (D), and if the pores contribute half the volume, ρ_b is half of ρ_S, namely 1.3-1.35 if ρ_S = 2.60-2.70. In sandy soils *d* may be as high as 1.6, whereas in aggregated loams and in clay soils, it can be as low as 1.1 gm/cm^3. The bulk density is affected by the structure of the soil, i.e., its looseness or degree of compaction, as well as by its swelling and shrinkage characteristics depending on clay content and wetness.

For the majority of field crops the optimal value of the bulk density of the soil (loam and clay) is within 1-1.2 gm/cm^3. Soil compaction reduces yields of crops. Growing oats on loamy soil, I.B. Revut and N.P. Poyasov (1953) noted that the growth *d* from 1.1 to 1.5 gm/cm^3 resulted in grain yield reduction of about 3.7 times (Table 37).

Table 37. Bulk Density Evaluation in Loamy and Clayey Soils (N.A. Kachinsky, 1965)

d gm/cm^3	Evaluation	d gm/cm^3	Evaluation
<1.0	Too loose or rich in OM	1.3–1.4	Very compacted plow layer
1.0–1.1	Typical for freshly plowed arable layer	1.4–1.6	Typical for plow layer
1.2	Slightly compacted plow layer	1.6–1.8	Very compacted illuvial horizons

True, that “Even in extremely compacted soil, however, the bulk density remains appreciably lower than the particle density, since the particles can never interlock perfectly and the soil remains a porous body, never completely impervious" (D. Hilled, 1982).

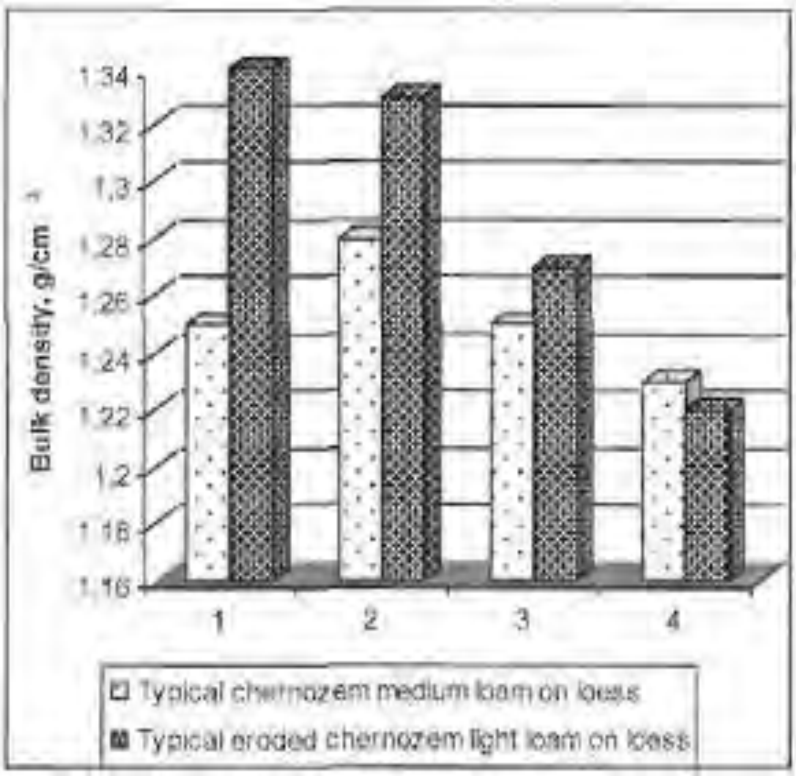


Fig. 84. Bulk density of Chernozems under different farming practices
(Iev. Berezhniak, I. Shevchenko, M. Berezhniak, 2013)

Porosity (*f* in American and P,% in Ukrainian designation) is an index of the relative pore volume in the soil. Its value generally lies within the range of 0.3-0.6 (30–60%). Ukrainian agropedologists prefer to compute soil porosity by the formula:

$$P_{\%} = (1-\frac{d}{D})\times 100 \qquad (27)$$

Coarse textured soils tend to be less porous than fine-textured soils, though the mean size of individual pores is greater in the former than in the latter. This explains why the rate of water infiltration in a sandy soil is greater than in a loamy one though the porosity of the former may be some 38-40% whereas that of the latter may reach 55% and more. Pore size distribution is very important, whatever the total porosity. N.A. Kachinsky (1965) proposed the following total porosity evaluation (Table 38). In various horizons of mineral soils soil porosity varies within a wide range of 25-80%. In horizons rich in humus (surface horizons) it usually is within 50–60% whereas in peat soils it may be 80–90%.

Table 38. Total Porosity Evaluation by N.A. Kachinsky

Total porosity in loamy and clayey soils during the period of crop growth, %	Evaluation	Total porosity in loamy and clayey soils during the period of crop growth, %	Evaluation
>70	Excessively loose (too porous)	<50	Unsatisfactory for a plow layer
65–55	Excellent for any cultured plow layer	40–25	Typical for compacted illuvial horizons, – excessively low
55–50	Satisfactory for a plow layer		

Depending on the size of the pores, the porosity may be capillary and noncapillary. Capillary pores are for water and noncapillary – for air. It is not bad if the ratio between the capillary and noncapillary porosities approaches a unity (an ideal agrophysical ratio).

It is important to know aeration porosity for any soil with field wetness. Aeration porosity at field capacity for any soil not connected with ground water and capillary fringe has a critical value of 15% which should not be lower. For peat soils this critical value is within 30–40%. If *W* designates soil wetness (% by oven dry weight), aeration porosity (AP,%) is computed according to the formula:

$$AP,\% = P\% - W \times d \tag{28}$$

Aeration porosity below 15% for mineral and 30% for organic soils favors the development of reduction processes. For agricultural soil physics it is important to know that the lower aeration porosity, the lower the diffusion coefficient of oxygen in soil air as compared to that in the open atmosphere.

18.2. Physico-Mechanical Characteristics of Soil

These characteristics include plasticity, cohesion (stickness), swelling, shrinkage, shear resistance, hardness (penetrometer resistance), and soil's resistance to tillage.

Plasticity is a property of a plastic soil. Plastic soil is "a soil capable of being molded or deformed continuously and permanently, by relatively moderate pressure, into various shapes" (H.D. Foth, 1990).

According to I.S. Kaurichev's textbook definition, "plasticity is the ability of the soil to change its shape under the impact of any external force without discontinuities and preserve this modified shape after the force ceases to act" (I.S. Kaurichev, 1996). To be in plastic state the soil has to possess certain wetness. The soil has plasticity within definite limits of wetness (Fig. 83). Largely due to the work of A.A. Attenberg and A. Cassagrande (1948), the Atterberg limits and related indices became very useful characteristics of assemblages of soil particles. Thus a soil is solid when dry, and upon the addition of water proceeds through the semisolid, plastic, and finally liquid states as shown in Fig. 85.

PI:	Soil Texture:
0	Sands
0–7	Loamy sands
7–17	Loams
>17	Clays

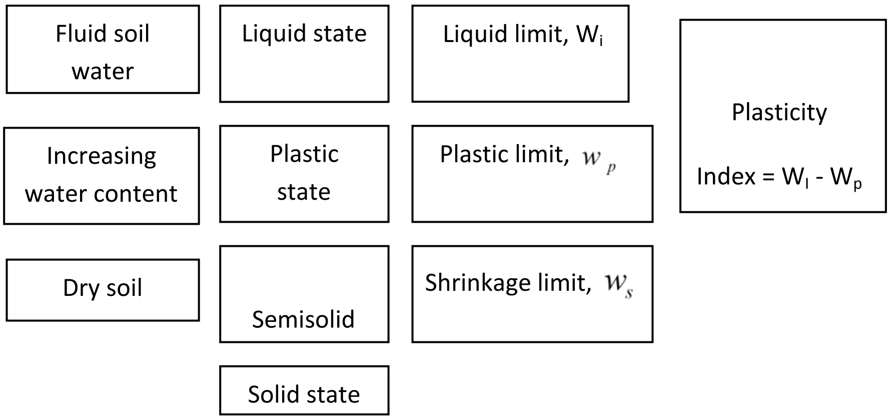


Fig 85. The Atterberg limits and related characteristics

Plasticity index (PI) is the difference between two soil moistures. The upper soil moisture is called liquid limit, the lower – plastic limit. Plasticity index expressed in percentage per weight of oven dry soil, as any soil moisture, is very much dependent on soil texture:

Plasticity Index also depends on the character of exchangeable cation in SAC Table 39.

Table 39. The Atterberg is Limits of Clay Minerals

Mineral	Exchangeable ion	Liquid limit, %	Plastic limit, %	Plasticity Index, %
Montmorilonite	Na	710	54	656
	Ca	510	81	429
	Mg	410	60	350
	Fe^{3+}	140	73	67
Kaolinite	Na	53	32	21
	Ca	38	27	11
	Mg	54	31	23
	Fe^{3+}	56	35	21

SOM (humus) decreases plasticity index of mineral soils. In spite of their serious limitations, there exist some simplified procedures for plasticity limits and plasticity index determination. The liquid limit is determined by measuring the water content and the number of blows (fallings) of a cup from a certain height required to close a specific width groove for a specific length in a standard liquid limit device. The plastic limit is determined by measuring the water content of the soil when wire of the soil 3 mm in diameter begins to crumble.

Soil Cohesion, or rather cohesiveness, is the ability of wet soil to stick to other bodies. As soil particles stick to the working organs of tillage implement the resistance to tillage increases and the quality of tillage operation degrades. The value of soil cohesiveness is measured by the force of a pull needed to detach a metallic plate (1 cm^3) from the surface of a sticky soil. Stickiness appears at soil wetness close to the plasticity limit. Soils with high OM (humus) content, like chernozems and meadow soils have no cohesiveness even at relatively high wetness (30-35%, Table 40).

Table 40. Cohesiveness and Penetromer Resistance of Some Soils from the Central Urals Area (N.A. Kachinsky)

Soil and land plot character	Horizon	Depth, cm	Wetness, %	Cohesiveness, gm/cm^2	Wetness, %	Penetrometer Resistance, kg/cm^2
Soddy-podzolic loam, Plowland, rye stubble	A	0–14	20.7	0	2.1	32-9
	A	0–14	36.2	2.63	29.5	1.9
Chernozem slightly sodicity at facted, light clay, plowland, spring wheat seedlings	A	0–14	41.3	0.13	8.9	9.8
	A	0–14	786	3.10	53.4	2.5
Sodic steppe soil, light clay, virgin land	A_1	0–8	30.7	0.26	6.2	87.3
	A_1	0–8	53.0	4.80	46.0	5.0

Soil cohesiveness is dependent on the composition of exchangeable cations linked with the SAC. Saturation with Ca^{2+} reduces cohesiveness, whereas exchangeable Na^+ may increase it very considerably. Heavy-textured soils have more “inclination” to be cohesive than the light-textured ones.

N.A. Kachinsky classifies the soils as excessively sticky(> $15gm/cm^3$), very sticky (5-15), sticky (2–5), and noncohesive (≤2 gm/cm^2). Cohesiveness may be useful too. Cohesiveness between clay particles is, in fact, the ultimate internal binding force within microaggregates. Organic polymers used to improve soil aggregation increase the cohesiveness of soil particles and thereby impart considerable stability to otherwise weak soil aggregates.

Swelling is soil volume increase or expansion in the process of wetting. Swelling is characteristic of fine-textured soils containing large amounts of colloidal matter. The reason why the soils swell in their imbibing of water molecules and their retention between the crystal unit layers of the layered silicates as well as retention of hydrated ions between these layers. Soil swelling is expressed in percentage by volume and calculated by the formula:

$$V_{sw} = \frac{V_1 - V_2}{V_2} \times 100 \qquad (29)$$

where V_{SW} is the percentage of soil swelling (increase of initial soil volume); V_1 is the volume of wet soil, and V_2 is the volume of dry soil.

The more plastic the soil, the more it swells. Montmorilonite- containing soils show a considerable swelling, whereas kaolinitic soils swell very little. Organic colloids also increase their volume at wetting. Exchangeable cations effect swelling very much, especially sodium which may cause the V_{SW} to reach 150%. Among the agronomically unfavorable consequence of swelling is the ruination of soil aggregates.

Shrinkage is the reduction of soil volume on drying. The extent of soil shrinkage is determined by the same agents as swelling. The greater the swelling, the more the soil shrinks. Soil shrinkage is, like swelling, expressed in % by volume and calculated by the formula:

$$V_{shr} = \frac{V_1 - V_2}{V_1} \times 100 \qquad (30)$$

where V_{shr} is the percentage of shrinkage of the initial soil volume; V_1 is the volume of wet soil; V_2 is the volume of dried soil.

Considerable shrinkage causes the formation of cracks in the soil mass which tear up roots of plants and increase physical evaporation of soil water.

Shear resistance (shear strength) is defined as the soil’s ability to resist an external effort trying to disunite particles composing soil mass. Interparticle friction and sliding are the causes of shear strength, as well as the forces of binding (cohesion) between the particles. Shear strength is determined by soil texture and microbiological composition as well as by the state of soil aggregation, wetness, and the character of its field use. Clayey soils possess the greatest shear strength, sandy ones – the lowest. Massive (nongranulated) soils when dry have the highest shear strength. Shear strength is expressed in kg/cm^2.

Soil hardness (penetrometer resistance) belongs to the so-called *in situ* properties of the soil, measured in the field. The most widely used field test method is penetration testing. The penetrometers are driven or pushed into the ground and the resistance to penetration is recorded. Some standard penetrometers have a working cross-section 1cm^2 in area. Penetrometer resistance is measured in kilograms of force (push) per 1 cm^2 of area. Penetration is made in the vertical direction. In soil mechanics the “standard penetration test” consists of driving the spoon into the ground by dropping a 63 kg mass from a height of 760 mm. The penetration resistance is reported in number of blows needed to drive the standard spoon 305 mm deep. Soil variability demands a great number of measurements (for example, 100 measurements per 1 m^2, using the Revyakin penetrometer in Ukraine). High penetrometer resistance is a sign of unfavorable physicochemical and agrophysical soil properties. Large amounts of energy (and fuel) are needed for soil tillage. Seeds find it hard to germinate and seedlings – to grow. Roots encounter resistance to penetration in the soil. Water infiltration and air components diffusion may also be too low. Soil hardness is very much dependent on its wetness. In dry soils (especially heavy textured ones) it increases rapidly.

Penetrometer resistance is also depend on the composition of exchangeable cations in the SAC. In chernozems saturated with Ca^{2+} it is 10-15 times lower than in solonetz (sodic) soil. High humus content also reduces soil hardness. By O.N. Sokolowski’s data, dried heavy clays had penetrometer resistance within 150-180 kg/cm^2. Plant roots meet a serious difficulty trying to penetrate the soil with hardness over 70-80 kg/cm^2. Soil resistance to tillage (e.g., plowing) is directly proportional to soil hardness.

Specific resistance to tillage is a technological characteristic of any cultivated soil. It is a force of plow pull divided by the area of a furrow slice. This area is the product of the depth of plowing and the width of it. Specific resistance to tillage (SRT) is expressed in kg/cm^2 and calculated by the formula:

$$SRT = \frac{F}{h \times b} \qquad (31)$$

where F is the force of plow pull (kg); h – the depth of plowing (cm), and b – its width (cm).

Soil texture and wetness are most determining factors affecting SRT. SRT should be determined when soil wetness is close to the soil’s physical maturity for tillage (PMT). For loams and clays this wetness corresponds to the minimal value of soil resistance to tillage and maximal ability of the soil to form granules (aggregates). Some coarse sand soils may have the lowest SRT when dry. But dry clays behave differently: their resistance to tillage reaches the highest value. Well granulated (aggregated) soils, other factors being equal, have a lower resistance to tillage compared with structureless (massive) soils. Table 41 gives some examples of SRT for a number of Eurasian soils.

Table 41. Soil's Specific Resistance to Tillage

Soil	Texture	Plot variant	SRT, kg/cm^2	Authors
Soddy-podzolic	Clay	Plowland	0.68	A.F. Pronin
The same	Heavy loam	the same	0.48	the same
» – » – »	Medium Loam	» – » – »	0.35	» – » – »
» – » – »	Light Loam	» – » – »	0.27	» – » – »
» – » – »	Loamy sand	» – » – »	0.18	» – » – »
Ordinary Chernozem	Clay	Virgin	0.7-0.8	I.B. Revut
The same	Loam	The same	0.6-0.8	the same
Soil	Texture	Plot variant	SRT, kg/cm^2	Authors
Sodic Chernozem	Clay	Plowland	0.82	» – » – »
Sodic soil (solonetz)	Clay	Virgin	1.21	D.I. Sarana
Syerozem (Grey desert soil)	Heavy loam	Irrigated Plowland	0.49	A.F. Pronin
» – » – »	Light Loam	» – » – »	0.34	» – » – »
» – » – »	» – » – »	Nonirrigated Plowland	0.27	» – » – »

Tillage is usually defined as the mechanical manipulation over the soil aimed at improving soil conditions affecting crop production. The practice of inverting the topsoil in order to bury manures and crop residues has become a less important function of tillage in modern field management. Crop residues can, and in many cases should, be left over the surface as a stable mulch to protect against evaporation and erosion. Tillage operation is especially consumptive of energy. The consumption of energy, as well as the wear and tear of tractors and implements, increases steeply as the depth of tillage increases. With the rising costs of fuel, the absolute and relative costs of tillage are almost certain to rise so that certain practices now in common practice may become prohibited.

Recent trends in tillage research are aimed at minizing tillage operations and travel. Both general physical and physico-mechanical soil characteristics have to be improved or at least preserved from any further degradation. Nonplow tillage to varying depth (depending on the crops in a crop rotation) favors the gradual disappearance of a plow sole. Regular applications of chemical amendments improve the soil in many ways, including soil structure. Perennial grasses also improve soil structure and reduce OM losses. Green manure crops and organic (animal) manures do the same.

Questions & Assignments

1. Name the general physical characteristics of soil.
2. Define each of the general physical characteristics of soil.
3. Define a formula in which all general physical characteristics of soil are interconnected.
4. Speak on agronomic evaluation of bulk density and total porosity.
5. Name physico-mechanical characteristics of soil.
6. In what cases and how should soil cohesiveness be reduced?
7. When kinds of swelling and shrinkage are in the way of crop growth? What practices can reduce them?
8. How can soil hardness be in the way of crop production?
9. Define specific soil resistance to tillage. What factors does it most considerably depend on?
10. Speak on the practices of soil resistance to tillage reduction.

Chapter 19

Soil Water. Water Related Properties and Regimes. Soil Water Management

Water is necessary for life. Soil always contains a certain amount of water. Soil wetness, or moisture, is determined in percentage of oven-dry (dried at 103 ºC) soil. It may also be expressed in percentage of soil volume, m^3/ha, and mm.

The main source of water in a nonirrigated soil is atmospheric precipitation. Annual precipitation averages 660 mm for the world's land surface. The soil located at the atmosphere-lithosphere interface plays an important role in determining the amount of precipitation that runs off the land and the amount that enters the soil for storage and future use. A plant may need from 200 to 1000 grams of water to synthesize 1 gram of dry matter. Elements of plant nutrition enter the plant with water.

The processes of weathering and soil formation, the activity of soil organisms depend very much on the moisture of the environment. All three physical states of water may be present in the soil: solid (ice), liquid, and gaseous (water vapor).

19.1. Categories of Soil Water and its Properties

The continuous nature of changes in the amount of soil water and corresponding changes in energy produce the soil water energy continuum. Liquid water and water vapor in the soil are subject to the action of various forces: sorption, capillary, osmotic, and gravitational. Under the action of these forces the properties of soil water, its mobility, and availability to plants change.

H.D. Foth (1990) considers in his textbook three forms of soil water: adhesion water, cohesion water, and gravitational water. In Ukrainian textbooks the picture is, without any reason, far more complicated. H.D. Foth (1990) sees the relationship between various forms of soil water as it is shown in Fig. 86. Adhesion water is strongly absorbed and very immobile and unavailable to plants. The gravitational water is beyond the sphere of cohesive forces, and gravity causes gravitational water to flow rapidly down and out of the soil (unless the flow is inhibited). The cohesion water is intermediate in properties and is the most important for plants growth.

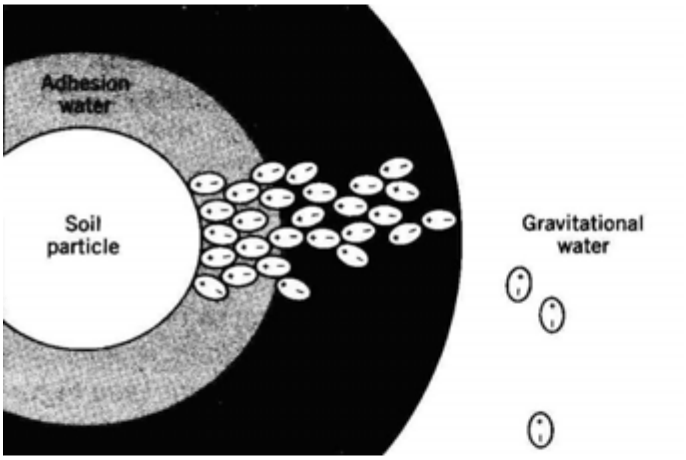


Fig. 86. Schematic drawing of the relationship between various forms of soil water (H.D. Foth, 1990)

The term *adhesion* water corresponds to the Ukrainian term *adsorption water* and strongly-held water of the films, cohesion water includes such categories of the Ukrainian system as loosely held water of the films and capillary water, whereas gravitational water corresponds to the Ukrainian category of the same name.

Adhesion water. Water molecules are strongly polar and attract each other through H bonding. Soil particles have sites that are both electrically negative and positive. The mutual attraction between water molecules and the surface of soil particles results in strong adhesive forces. “If a drop of water falls onto some oven dry soil, the water molecules encountering soil particles will be strongly adsorbed and the molecules will spread themselves over the surfaces of the soil particles to form a thin film of water. This layer, or film of water, which will be several water molecules thick, is called *adhesion wate*r” (H.D. Foth, 1990). Adhesion water is immobile, has the lowest energy level, and is generally unavailable for use by plant roots and microorganisms.

Cohesion water. Beyond the sphere of strong attraction of soil particles to water molecules of the adhesive water layer, cohesive forces operate between water molecules because of hydrogen bonding. “Water molecules attach themselves to the adhesion water, resulting in an increase in the thickness of the water film. Water that is retained in soils because of cohesive forces is called *cohesion water*” (H.D. Foth, 1990).

About 15 to 20 molecular layers is the maximum amount of water that is normally absorbed to soil particle surfaces. In Ukraine it is known as maximum adsorption capacity (MAC). It is about 0.6-0.7 of the so-called maximum hygroscopic moisture acquired by the air-dry soil samples kept from one to three weeks at relative air humidity close to saturation $(P/P_0 = 1)$, and actually being equal to 0.96-0.98 till the soil reaches constant weight. With increasing distance from the soil particle surface toward the outer edge of the water film, the strength of the forces that hold the water decreases. (MAC) and the energy and mobility of the water increase. Compared with adhesion water which is essentially immobile and unavailable to plants, the cohesion water is slightly mobile and generally available to plants. There is water-energy gradient across the water films and the outermost water molecules have the greatest energy, the greatest tendency to move, and the greatest availability for plant roots.

As roots absorb water over time, the soil water content decreases, and water films become thinner. Roots encounter water with decreasing energy content or mobility (availability). At some points, water may move so slowly to roots that plants wilt because of a lack of water. If the plants fail to recover in humid atmosphere of a special chamber the soil is at the permanent wilting point (PWP). In practical agronomy the PWP is often calculated according to the formula:

$$PWP(\%) = MHM \times 1.5 \qquad (32)$$

where MHM is the maximum hygroscopic moisture of the soil.

In order to contain water available to plants the soil moisture must exceed the PWP. When soil microspores (capillary pores) retain all possible water against gravity, the soil is considered to be at field capacity (FC). “Field capacity, in situ (field water capacity) is the content of water, on a mass or volume basis, remaining in a soil 2 or 3 days after having been wetted with water and after free drainage is negligible” (H.D. Foth, 1990).

Peters (D. Hillel, 1982) wrote about FC determination: “Wet the soil to the depth of interest. Allow the soil to drain for approximately 2 days”. There are many drawbacks in such a definition but the characteristic itself is, if not theoretically, practically important. One must decide at the outset what drainage rate one is willing to consider negligible. Soil matric suction at field capacity is approximately 1/3 bars, whereas at the PWP it is about 15 bars.

The aeration or macrospores at FC are air filled. A typical good loamy surface soil at FC has a volume composition of about 50% solids, 25% water, and 25% air.

Soil capillaries also suck the water from the ground water table, forming the so-called water-saturated zone of capillary fringe over it. Soil wetness within a capillary fringe is conventionally called capillary capacity. Near the water table this wetness is close to full capacity of soil moisture, when practically all pores are tilled with water. At the upper edge of a capillary fringe soil moisture is close to FC.

Gravitational water. An example may be ground water, which is that portion of the water below the surface of the ground at a pressure equal to or greater a atmospheric. In soils with clay pan, a very slowly permeable subsoil layer permits little water, in excess of field capacity, to move downward and out of the soil.

“Soil water that exists in certain pores, and that is normally removed by drainage because of the force of gravity, is gravitational water” (H.D. Foth, 1990).

Gravitational water in soil is detrimental when it creates oxygen deficiency. Gravitational water is not considered available to plants because it normally drains out of soils within a day or two after the soil becomes very wet. When water filled macrospores are allowed to drain, the water in the center of the macrospore has the most energy and is the most mobile. This water leaves the pore first and at the fastest speed. This is followed by other water molecules closer to the pore edges until all of the gravitational water has drained, leaving the cohesion and adhesion water a film around the periphery of the macropore.

Water Vapor. When soils are unsaturated, some of the pore space is air filled and water moves through the pore space as vapor. The driving force is the difference in vapor pressure (in water potentials). Generally, vapor pressure is high in warm and moist soil and low in cold and dry soil. Vapor flow occurs from warm and moist soil to and into cold and dry soil. In summer, vapor flow travels from the warmer upper soil horizons to the deeper and colder soil horizons. In cold season, vapor flow tends to be upward. It is encouraged by low air temperature. Upward vapor movement, and condensation onto the surface of the soil are often felt in the late night and early morning hours.

Vapor movement is minor in most soils, but it becomes important when soil cracks appear in the surface horizons and water is lost from the soil from the surfaces of the cracks.

Capillary forces and surface tension are very important for the behavior of cohesion water in the soil. The movement of water in soil capillaries is caused by the difference in capillary pressures arising as a result of different meniscus curvatures.

Osmotic forces are involved in adsorption of water molecules by ions (hydration) from the dissolution of soluble salt. The forces that affect the energy level of soil water, and the mobility and availability of water to plants are adhesion, cohesion, gravity, and osmotic forces. As dry soil wets over time, the soil water increases in energy, mobility, and availability for plants. As wet soil dries over time, the energy, mobility and plant availability of the remaining water decreases.

There is a relationship between the energy of water and pressure of water. It is much easier to determine the pressure. The greater the water pressure, the greater the tendency of water to move and do work, and the greater the energy of water.

19.2. Soil Water Potential

In water-saturated soil, the water pressure is zero at the surface of the water table and the water pressure increases with increasing soil depth.

Water in unsaturated soil has a negative pressure or is under tension. The water pressure in unsaturated soil decreases with increasing distance above a free water surface or water table.

The pressure of soil water is expressed by the soil water potential. Technically it is the amount of work that must be done per unit quantity of water to transport or move reversibly and isothermally an infinitesimal quantity of water from a pool of pure water at specified elevation and at atmospheric pressure to the soil water at the point under consideration. So it is the amount of work needed to move water from a reference pool to another point. The symbol for it is psi, Ψ. The total water potential Ψ_t is made of several subpotentials, including the gravitational (Ψ_g), matric (Ψ_m) and osmotic (Ψ_0):

$$\Psi_t = \Psi_m + \Psi_g + \Psi_0 + ... \quad (33)$$

Not all of the separate potentials act in the same way. Their separate gradients may not always be equally effective in causing flow. For example, the osmotic potential gradient requires a semipermeable membrane to induce liquid flow.

What was the main reason to employ the concept of soil water potential? It provides a unified measure by which the state of water can be evaluated at any time and everywhere within the soil-plant-atmosphere continuum.

19.2.1. Gravitational Potential

Every body on the earth's surface is attracted toward the earth's center by a gravitational force equal to the weight of the body, the weight being the product of the mass of the body and the gravitational acceleration. To raise a body against this attraction, work must be expended. This work is stored by the raised body in the form of **gravitational potential energy**.

The gravitational potential Ψ_g is determined by the position of water in the gravitational force field. This potential is very important in water-saturated soils. It accounts for the movement of water from high to low elevation or through saturated soils. The higher the water above the reference point, the greater the gravitational water potential. For the sake of convenience, it is customary to set the reference level at the elevation of a pertinent point within the soil, or below the soil profile being considered. In such cases the gravitational potential can always be taken as positive or zero. If the soil surface is chosen as the reference level, as is often done, the gravitational potential for all points below the surface is negative with respect to that reference level.

At a height Z above a reference, the gravitational potential energy E_g of a mass M of water occupying a volume V is

$$E_g = MgZ = \rho_W VgZ \tag{34}$$

where ρ_W is the density of water and g the acceleration of gravity.

In terms of the potential energy per unit mass, the gravitational potential is

$$\phi_g = gz \tag{35}$$

and in terms of potential energy per unite volume is

$$\phi\ ,_v = \rho_W gZ \tag{36}$$

The gravitational potential does not independ on a chemical and pressure conditions of water, but depend only on relative evaluation.

The water pressure or gravitational potential at the top of water-saturated soil decreases as the water level drops. At the drainage tubing (the reference level) the gravitational potential is zero.

19.2.2. Matric Potential

The matric potential is also called the pressure potential, but it expresses negative units. The *pressure potential* is often termed *capillary potential*. *Matric potential* is a more recent term. Unsaturated soil contains no gravitational water.

The important forces affecting water movement are adhesion and cohesion. This potential of soil water results from the capillary and adsorptive forces due to the soil matrix. These forces attract and bind water in the soil and lower its potential energy below that of bulk water. Capillarity results from the surface tension of water and its contact angle with the solid particles. In an unsaturated (three-phase) soil system, curved menisci form which obey the equation of capillarity

$$P_0 - P_C = \Delta P = \delta(\frac{1}{R_1} + \frac{1}{R_2}) \tag{37}$$

where P_0 is the atmospheric pressure, conventionally taken as zero, P_C the pressure of soil water, which can be smaller than atmospheric, Δ **P** is the pressure deficit, or subpressure, of soil water, δ is the surface tension of water, and R_1 and R_2 are the principal radii of curvature at a point on the meniscus.

The soil is not like a bundle of capillary tubes. In addition to the capillarity phenomena, the soil also exhibits adsorption, which forms hydration envelopes over the particle surfaces (Fig. 87).

The presence of water in films as well as under concave menisci is most important in clayey soil and at high suctions. In sandy soils adsorption is relatively unimportant and the capillary effect predominates. When the rain falls on dry soil, the water molecules from a pool of pure water move to the water films on soil particles.

No energy input is required for the movement of water in raindrops to water films around dry soil particles. In fact, a release of energy occurs resulting in a decrease in the water potential. Adsorbed water has lower potential to do work than water raindrops. In other words, negative work is required to transfer water from raindrops to the films on soil particles. That is why the matric potential has a negative sign.

The matric potential varies with the content of soil water. The drier a soil is, the greater is its tendency to wet and the greater is the release of energy when it becomes wetted. The release of heat is called *the heat of wetting.*

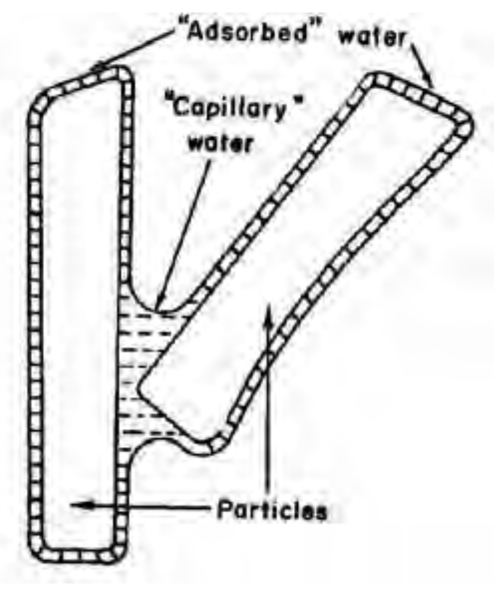


Fig 87. Water in unsaturated soil subject to capillarity and adsorption

Some soil physicists prefer to separate the positive pressure potential from the matric potential, assuming the two to be mutually exclusive. Soil water may exhibit either of the two potentials, but not both simultaneously. An additional factor which may affect the pressure of soil water is a possible change in the pressure of the ambient air. In general, this effect is negligible. In the absence of solutes, the liquid and vapor phases in an unsaturated porous medium are related by formula:

$$\phi_m = RT\ln(P/P_0) \tag{38}$$

where P/P_0 is the relative humidity, R is the gas constant for water vapor, and T is the absolute temperature. The term P/P_0 represents the ratio of the partial pressure of vapor in the air phase relative to the partial pressure at saturation.

19.2.3. Osmotic Potential

The presence of solutes in soil water causes the osmotic effects. Example of this is the following solutes lower the vapor pressure of soil water. This phenomenon may not effect liquid flow in the soil significantly, but osmotic effect is important in the interaction between plants roots and soil.

The expression *osmotic pressure of a solution* can be misleading. What a solution exhibits relative to the pure solvent (say, water) is not an excess pressure. On the contrary, a suction such that will draw water from a reservoir, of pure water brought into contact with the solution across semipermeable boundary. The term *semipermeable membrane* is hallowed by tradition since Van't Hoff who coined it. Today, only its educative convenience remains. Actually, it is something of a misnomer. What is really important is that in dilute solutions, the osmotic pressure is proportional to the concentration of the solution and to its temperature according to the following equation:

$$\Pi = MRT \tag{39}$$

where Π is the osmotic pressure in atmospheres, M is the total molar concentration of solute particles (whether molecules or dissociated ions), T is the temperature in degrees Kelvin, and R is the gas constant

$(0.08205\frac{liter \times atm.}{\deg \times mole})$

In soils containing a large amount of soluble salt (saline soils) the osmotic potential has the effect of reducing water uptake by roots, seeds, and microorganisms. The osmotic potential has little, if any, effect on water movement within the soil, since diffusion of the hydrated ions creates an uniform distribution and the hydrated ions do not contribute to a differential in metric potential in one region of the soil as compared to the other.

19.2.4. Quantitative Expression of Soil-Water Potential. Soil-Moisture Characteristic Curve

Physically, the soil-water potential may be expressed in energy per unit mass, the dimensions being L^2T^{-2}; energy per unit volume ($ML^{-1}T^{-2}$). This method is convenient for expreming osmotic and pressure potentials, but is seldom used for the gravitational potential and energy per unit weight (hydraulic head). Whatever can be expressed in units of hydroscopic pressure can also be expressed in the height of a liquid column corresponding to the given pressure. A pressure of 1 atm is equivalent to a vertical water column, or hydraulic head, of 1033 cm, and to a mercury head of 76 cm. This last method of expression is simple and often more convenient. Instead of

$$\phi = \phi_g + \phi_p \quad (40)$$

one could write

$$H = H_g + H_p \quad (41)$$

which reads: the total potential head of soil water (H) is the sum of the gravitational (H_g) and pressure (H_p) potential heads. H is called *the hydraulic head.*

On attempting to express the negative pressure potential of soil water (relative to atmospheric pressure) in terms of an equivalent hydraulic head, we must contend with the fact that this head may be of the order of 10,000 or even 100,000 cm of water. To avoid the use of such cumbersomely large numbers, J. Schofield (1935) suggested the use of pF, which he defined as the logarithm of the negative pressure (tension, or suction) head in centimeters of water. A pH of 3 is the tension head of 1000 cm H_2O If ϕ is the potential in terms of energy per mass, p – the potential in terms of pressure, and H – the potential head, then

$$\phi = \frac{p}{\rho_W} \tag{42}$$

$$H = \frac{p}{\rho_{W g}} = \frac{\phi}{g} \tag{43}$$

In a saturated soil at equilibrium with free water at the same elevation, the actual pressure is atmospheric, and hence the hydrostatic pressure and the suction (or tensions) equal zero.

As more and more water is drawn out of the soil, more and more of relatively large pores which cannot retain water against the suction applied will empty out.

Considering the capillary equation (assuming that soil pores are approximately cylindrical), we can readily predict that a gradual increase in suction will result in emptying progressively smaller pores until, at high suction values, only very narrow pores retain water.

$$-P = \frac{2\gamma}{r} \tag{44}$$

An increase in soil-water suction is associated with a decreasing thickness of the hydration envelopes covering the soil-particle surfaces. Increasing suction is thus associated with decreasing soil wetness.

Childs (1940) paid attention to the fact that the amount of water remaining in the soil at equilibrium is a function of the matric suction. This function is usually measured experimentally and represented graphically by a curve known as the soil-moisture characteristic curve $(SMCC)$*.

The amount of water retained at relatively low values of matric suction (say, between 0 and 1 bar of suction) depend primarily upon the capillary effect and the pore-size distribution, and hence is strongly affected by the structure of the soil. Water retention in the higher suction range is increasingly due to absorption and is thus less influenced by the structure and more by the texture and specific surface of the soil material. The water content at a suction of 15 bar (PWP) is fairly well correlated with the surface area of a soil and would represent, roughly about 10 molecules layers of water if it were distributed uniformly over the particle surface. So, the $SMCC$ is strongly affected by soil texture (Fig. 88).

---

* $SMCC$ has its Ukrainian equivalent as follows: “основна гідрофізична характеристика ґрунту”.

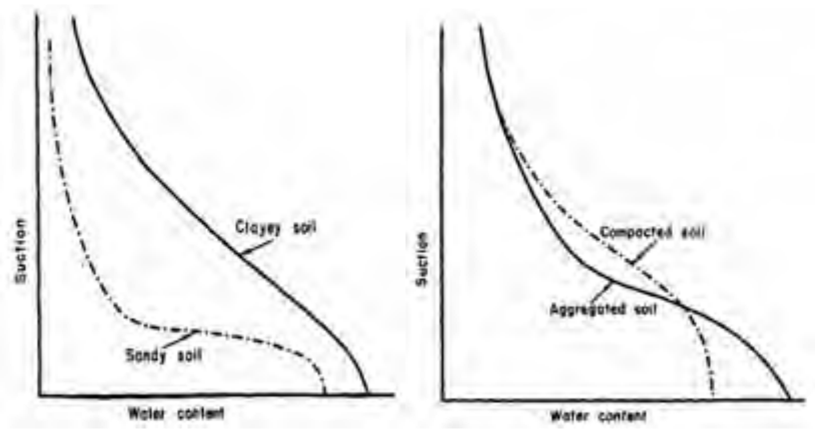


Fig 88. The effect of soil texture and structure on soil-water retention.

In a nonshrinking soil, the *SMCC*, once obtained, allows calculation of the effective pore-size distribution (i.e., the volumes of different classes of pore size). If an increase in matric suction from Ψ_1 to Ψ_2 results in the release of a certain volume of water, then that volume is evidently equal to the volume of pores having a range of effective radii between r_1 and r_2, where Ψ_1 and r_1, and Ψ_2 r_2 are related by the equation of capillarity, namely :

$$\Psi = \frac{2\gamma}{r}. \qquad (45)$$

The matric potential can be measured in several ways. Vacuum gauge potentiometers consist of a rigid plastic tube having a porous fired clay cup on one end and vacuum gauge on the other. The potentiometer is filled with pure water, and at this point the vacuum gauge reads zero. The potentiometer is then buried in the soil so that the porous cup has good contact with the surrounding soil (Fig. 89).

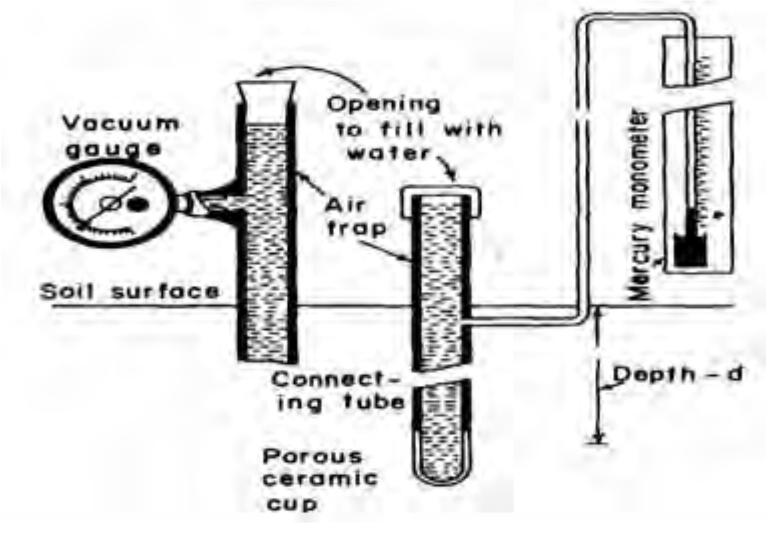
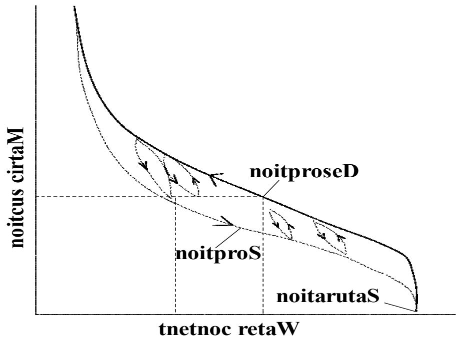


Fig 89. Essential parts of a tensiometer (D. Hillel, 1982).

The relation between matric potential and soil wetness can be obtained in two ways: 1) **in desorption,** by taking an initially saturated sample and applying increasing suction to gradually dry the soil while taking successive measurement of wetness versus suction; and 2) **in sorption**, by gradually wetting up an initially dry soil sample while reducing the suction. Each of the two procedures yields a continuous curve, but the two curves will not coincide. The equilibrium soil wetness at a given suction is greater in desorption (drying) than in sorption (wetting). This phenomenon is called *hysteresis* (Fig. 90).



The hysteresis may be attributed to several causes, the main of them being the geometric nonuniformity of the individual pores (which are generally irregularly shaped voids interconnected by smaller passages), resulting in the “inkbottle” effect.

Fig 90. The suction-water content curves in sorption and desorption (D. Hillel, 1982).

19.3. Soil Water Movement

The driving force of water movement is the difference in water potentials between two points. The water potential gradient (f) is the water potential difference between two points divided by the distance between the two points. The rate of water flow is directly related to the water potential gradient.

The velocity of water flow is also affected by the soil’s ability to transfer water or **the hydrolytic conductivity** (K). Water flow through large pores is faster then through small pores. Flow is faster when conductivity is greater. The velocity of flow, V , is equal to the water potential gradient times the hydrolytic conductivity, K :

$$V = Kf \qquad (46)$$

The hydrolytic conductivity is the ability of the soil to transmit water. It is determined by the nature and size of soil pores through which the water moves.

Conductivity is closely related to pores size. The larger pores give the sandy soils a greater ability to transmit water as compared to clayey soils.

Water Movement in Saturated Soil. In water-saturated soils, water moves much more rapidly, as all soil pores are water filled and water moves through the largest pores. As the water content of soil decreases, water moves through pores with decreasing size and, therefore, more and more slowly. K decreases with decreasing soil water content.

The potential gradient (f) between two points in water-saturated soil is the head (h). The gradient is inversely related to the distance of flow through the soil, d:

$$f = \frac{h}{d} \tag{47}$$

We can now write

$$V = K \times \frac{h}{d} \tag{48}$$

The quantity of water, Q_W, that will flow through a soil (or a pipe) depends on the cross sectional area, A, and the time of flow, t **t**:

$$Q_W = K\frac{h}{d}At \tag{49}$$

This equation is named after H. Darcy (Fig 91).

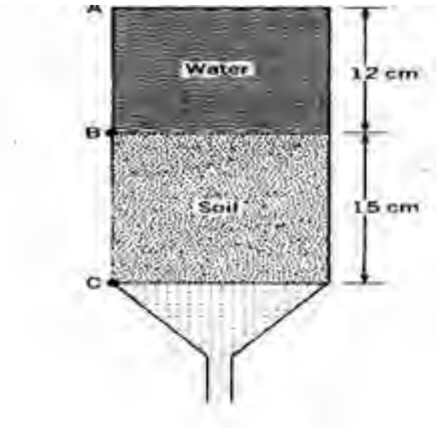


Fig. 91. A setup similar to that which H. Darcy used to study water movement in porous materials ($A = 100cm^2, t = 1hr$) (D. Hillel, 1982).

For the setup depicted in Fig 10.6, the above stated equation is “calculated” as follows:

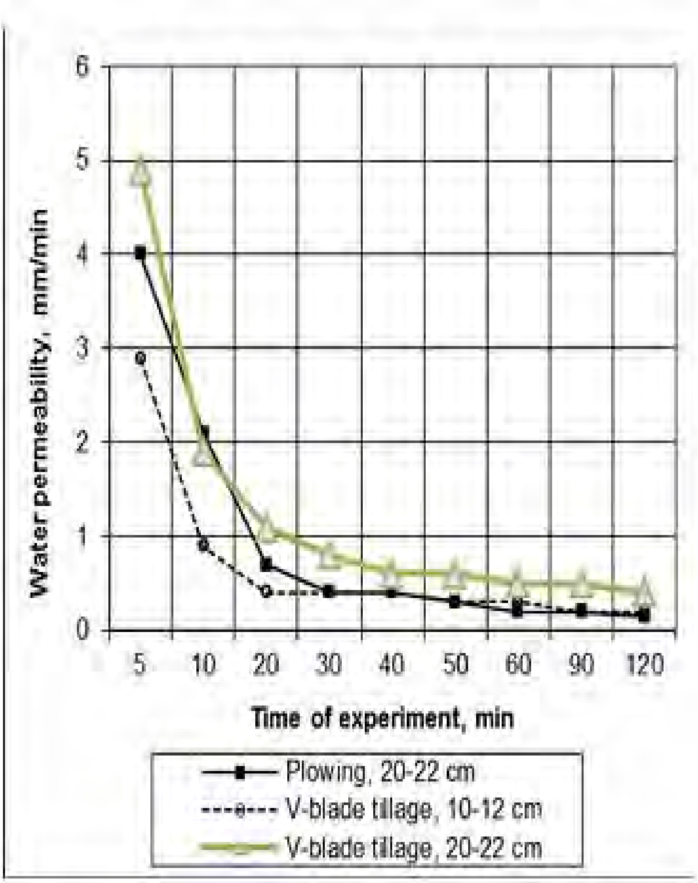
$$360 \quad cm^3 = K \times \frac{27 \quad cm}{15 \quad cm} \times 100 \quad cm^3 \times 1hr \quad then\, K = 2 \quad cm/hr \tag{50}$$

A sand soil could have a saturated hydrolytic conductivity that is several hundred times greater than that of clay soil. If the distance of water movement through soil is doubled, the quantity of water flow is halved. The further soil is from the drain lines, the longer it will take for water to exit the soil.

Water Movement in Unsaturated Soil. As drainage proceeds over time, water movement shifts to smaller and smaller pores. The result is a rapid decrease in both hydrolytic conductivity and in the rate of drainage. Field soils with average saturated hydraulic conductivity will, essentially, stop draining within a day or two after through complete wetting. The gravitation water exits the soil. The soil retains the maximum amount of water against the force of gravity. Such soil is at **field capacity.** Its metric potential will be within – 0.1-0.3 bars. Reaching field capacity (FC) means that the hydraulic conductivity becomes drastically reduced, say, about 1% of as great as that at saturation. Gravity is always operating, but after FC is attained, however, matric forces control water movement. The flow tends to be horizontal from soil to roots, except for some vertical movement to the soil surface, resulting from loss of water by evaporation.

Unsaturated hydraulic conductivity decreases rapidly when plants absorb water and soils become drier than field capacity. Somewhere midway between FC and the PWP hydraulic conductivity rapidly decreases, which is a result of **capillary discontinuity**. Stratified soils have layers, or horizons, with different physical properties, resulting in differences in hydraulic conductivity. The downward movement of water may be altered when the downward moving water front encounters a layer with a different texture. Matric forces (cohesion and adhesion) control water movement; water tends to move upward almost as rapidly as downward.

If the water is moving through loam soil and the wetting front encounters a dry sand layer, the wetting front stops its downward movement. The wetting front then continues to move laterally above the sand layer, because the unsaturated hydraulic conductivity in the dry sand is less than that of the wet silt loam soil above. As the soil above the sand layer becomes wetter over time, the water potential gradient between the moist soil and dry soil increases. In time, the water breaks through the silt loam soil and enters the sand layer. The low unsaturated hydraulic conductivity of any sand and gravel layers underlying loamy soils enables the loam soil layer to retain more water than if the soil had a loam texture throughout.



If a clay layer underlies a loamy soil, the wetting front will immediately enter the clay. However, the clay's capacity to transmit water is very low by comparison, and water will continue to move downward in the loam soil at a faster rate than through the clay layer. Under these conditions, the soil saturates above the clay layer as it does when a clay pan exists in a soil. So, it is apparent that well-drained soils tend never to saturate unless some underling layer interrupts the downward movement of a wetting front and soil saturates from the bottom upward.

Fig. 92. Soil permeability of typical eroded Chernozem under different tillage, alfalfa 2nd year (E. Berezhniak, 2013)

At the junction of soil layers, or horizons, of different texture, downward moving water tends to build-up and accumulate. This causes a temporary increase in the availability of water and nutrients, unless very wet and saturated soil is produced, and results in increased root growth and uptake of nutrients and water near or at these junction.

19.4. Soil Properties Related to Water

There still exists a tradition in Ukrainian pedology to classify these properties into: 1) water-holding capacity, 2) soil infiltrability to water (infiltration rate), and 3) water-elevating soil capacity (the height of capillary rise).

Water-holding capacity manifests itself in various forms and is defined as the ability of soil to retain water by the action of sorption and capillary forces (adhesion and cohesion forces in a modern concept).

Thus, the **maximal adsorbing capacity** (*MAC*) is the maximum amount of water held by a given soil by the forces of adhesion. **Field capacity** (*FC*) is another and already defined form of the water holding capacity. **Maximum hygroscopicity** (*MH*) (hygroscopic moisture) has been defined too, as well as the PWP which can be approximately calculated as $1.5\times MH$. PWP is sometimes called the wilting coefficient. **Full capacity** corresponds to soil moisture at which when all soil pores are filled with water (saturation capacity may serveses its synonym, SC). SC may be approximately calculated for nonswelling soils by dividing total porosity by the bulk density of the soil. For mineral soils it can be computed by one of our regression equations: $SC = 100 / d - 5d - 30$, where d is the bulk density of a soil. Moisture of capillary discontinuity (MCD) is approximately 0.65-0.70 of the FC . Capillary capacity (CC) is a term still used in Ukraine. Conventionally, it amounts to soil wetness in the lower part of a capillary fringe just some dozen centimeters above the ground water table. In the upper edge of the capillary fringe, soil moisture is close to the FC.

Infiltration Rate, mm/hr:	Grade:
> 1000	Rather rapid
1000–500	Excessive
500-100	The best
100-70	Good
70-30	Satisfactory
<30	Unsatisfactory

Infiltration rate is the stable rate of water infiltration in the soil after percolation reaches its stable minimum. N.A. Kachinsky recommended the following estimation scale for the infiltration rate at hydraulic head equal to 5 cm and the temperature of water – to 10ºC:

Height of Capillary Rise (HCR) depends on soil texture. For sandy soils it may be within 0.5-0.7 m, whereas for (HCR) loams it may reach 3-6 m. A good condition of soil structure reduces the HCR.

19.5. Availability of Soil Water to Plants

Plants use large quantities of water. For wheat to produce the yield of grain of 5 mt/ha it is necessary to absorb 500 mm (5,000 mt/ha) of water during the period of plant growth (data for the Forest-Steppe zone of Ukraine).

The water-supplying power of soil is related to the amount of available water a soil can hold. The available water is the difference in the amount of water at FC (30 kPa or -0.3 bar) and the amount of water at the PWP (-1,5000kPa or -15 bar). In Ukraine characteristic is called *the active moisture range in the soil* and is expressed primarily in mm or t/ha. Assuming the PWP to be equal to $1.5MH$ (maximal hygroscopicity), this range can be computed for a given elementary layer by the formula:

$$AMR\ (active\ moisture\ range), mm = 0.1(FC - 1.5MH) \times h\ \times\ d_v \qquad (51)$$

where h is the depth of a layer in cm and d_V is the bulk density of the soil in a given layer. AMR exceeding 160 mm is evaluated as excellent, whereas that below 60 mm – as unsatisfactory in one meter layer of soil.

The amount of available water is related to both texture and structure. It depends on the nature of surfaces and pores, or soil matrix. Soils high in silt (silt loams, like the major part of Ukrainian loess soils) tend to have the most optimum combination of surfaces and pores. They have the largest available water-holding capacity. When such soil is at the PWP, some 50 mm of rain or irrigation, which infiltrates completely, will bring the soil to FC to a depth of about 30 cm. Sands have the smallest available water-holding capacity: they are droughty soils. Most of available water in sands is held at relatively high water potential, compared with loams and clays, so that much of the plant available water in sandy soils can move rapidly to roots when water near roots is depleted, because of the relatively high hydrolytic conductivity at these relatively high potentials (H.D. Foth, 1990). Plants growing on clay soils, by contrast, may not absorb as much water as they could use during the day because of lower hydrolytic conductivity. This results in the tendency to conserve the available water in fine-textured soils. A serious water stress may appear in the soil at the moisture of capillary discontinuity when the bulbs of air appear in the smaller pores of soil. Availability of soil water to plants drastically reduces. Only desert plants can thrive with very low water potentials. Some plants shed their leaves during periods of extreme water stress.

Good management should maintain soil moisture within the range of FC to the MCD (moisture of capillary discontinuity) during the growing season. Crops that have a long growing season and a deep root system, like alfalfa, absorb a greater proportion of their water below 30 centimeters.

Most plants cannot tolerate the low oxygen levels of water-saturated soils. Water saturation of soil may kill many kinds of plants. These plants experience an increase in plant growth from saturation to near field capacity. As soils dry beyond field capacity, increased temporary wilting during the daytime causes a reduction in photosynthesis. Plant growth tends to be at a maximum when growing in soil near FC. The integrated supply of both oxygen and water is the most favorable.

19.6. Soil Water Regime

Soil water regime may be defined **as the set of phenomena of water input into the soil, its movement and retention in soil horizons, and its losses from the soil.** Quantitatively soil water regime may be expressed by the equation of soil water balance. The contributing sources of soil water are in the left side of the equation and the kinds of water losses from the soil – in its right side:

$$\begin{aligned}&B_0+B_{pr}+B_{gm}+B_{cond}+B_{s.\inf.}+B_{l.\inf.}=\\&B_1+B_{evap.}+B_{tr.}+B_{\inf ilt.}+B_{s.runoff}+B_{l.runoff}\end{aligned}\qquad(52)$$

In this equation B_0 and B_1 are the amounts of water in a definite soil layer (taken for research) at the beginning and at the end of the period of observations; B_{pr} is the amount of water coming to soil with precipitation; B_{gw} is the amount of water reaching the layer of the soil by capillary rise from the ground water; B_{cond} is the amount of water formed from the condensation of vapor; $B_{s.\inf.}$ is the amount of water coming into the soil as the surface inflow; $B_{l.\inf.}$ is the amount of water coming into the soil from the lateral subsoil or sub horizon inflow, $B_{evap.}$. is the amount of water evaporated from the soil surface; B_{tr} is the amount of water lost from the soil by transpiration; $B_{s.runoff}$ is the amount of water lost from the soil by surface runoff (or rather never reaching the soil because of it); and $B_{1.runoff}$ is the amount of water lost from the soil by the lateral (subsurface) runoff.

Very often such an equation is composed for the cyclical period of one year. In many concrete cases by far not all the sources of water addition to and removal from the pool of soil water are actually present in reality and the equation becomes much simpler. For some level land plots it may acquire the form:

$$B_{pr}+B_{gm}+B_{s.\inf.}=B_{evap.}+B_{tr.}+B_{s.runoff}\qquad(53)$$

Metric system allows to compute the amount of water (B in any elementary soil horizon, h cm deep in mt/ha (m^3/ha) according to the simple formula:

$$B = W \times H \times d_V \,, \qquad (54)$$

where W is soil wetness in a given layer and d_V is the bulk density (gm/cm^3) of the soil in a given layer. Multiplying B by 0.1 allows to express it in millimeters. The amount of water in 1-meter layer is computed by summing up the amounts in the elementary layers, the total thickness of which is 1 meter.

The amount of available water in a layer h cm thick, with bulk density of the soil of d_V, and field wetness of W, is calculated (in mt/ha) by the formula:

$$B = (W - PWP) \times h \times d_V \qquad (55)$$

B in mm is obtained by the formula:

$$B, mm = 0.1(W - PWP) \times h \times d_V \qquad (56)$$

The amount of available water in the soil is evaluated according to the following scale (Table 42).

Table 42. Available Water Amount Evaluation

Soil Layer Thickness, cm	Amount of Available Water, mm	Evaluation Grade
0–20	> 40	Good
	40–20	Satisfactory
	<20	Unsatisfactory
0–100	>160	Very good
	160–130	Good
	130–90	Satisfactory
	90–60	Unsatisfactory
	< 60	Very unsatisfactory

19.6.1. Types of Soil Water Regime

Water balance may acquire different forms in various soil-climatic zones and areas. In Ukraine the types of soil regimes are identified using basically two criteria:

1) the ration between the annual amount of rainfall and annual evapotranspiration or evaporation from an open water area (humidity coefficient), and 2) the depth of ground water table. A.A. Rode identifies 6 types of soil water regime:

1. **Permafrostic**, characteristic of the areas with permafrost. The mean annual soil temperature is below 0ºC, the depth of freezing in winter may exceed the depth of thawing in summer. The surface layer that thaws in summer and freezes in winter is the active layer. The active layer is used by plants. The base of the active layer is the upper surface of the permafrost, or the permafrost table. The melting of frozen soil ice in summer results in wet soil conditions, even if there is little or no rainfall.

2. **Percolative** (humidity coefficient, hc>1), which may be widespread in localities with annual rainfall exceeding the evaporability (evaporation of water from an open water surface). Downward flow of water dominates, and the soil and subsoil are percocolated by water to the depth of ground water table. The soil profile is leached of carbonates and other soluble products of soil formation.

Podzolic soils of taiga were formed under the impact of this type of water regime, as well as soddy-podzolic soils, oxisols (redzems) and yellowzems of the humid subtropics.

Stagnant subtype of this type of soil water regime develops on areas with high water table or on impermeable subsoils. Percolate type of soil water regime favors the development of soil acidity;

3. **Periodically percolative type** (hc = 1, being within 0.8-1.2). The soil profile is washed off by percolating water once in a period of several years. The soils are mainly leached and acid. The examples may include grey forest soils, podzolized chernozems, and leached chernozems (Mollisols).

4. **Nonpercolative** ($h_c < 1$) type of soil water regime is typical of localities where the precipitation water wets only the surface soil and subsoil horizons never reaching the ground water. After the rain, some water will evaporate from the soil surface. This creates a water potential gradient between the surface of the drying soil and underlying moist soil, and some water in the underlying moist soil will migrate upward. This phase of soil drying is characterized by rapid water loss. As drying near the soil surface occurs, a rapid decrease in hydraulic conductivity occurs at the soil surface. Water migration upward is greatly reduced and the evaporative demand at the surface greatly exceeds the water movement upward from moist soil on a hot summer day. As a result, the immediate surface of the soil becomes air dry within a few hours and the hydraulic conductivity for liquid water approaches zero. This phase is associated with a sharp decline in water loss on a fallow field. When

another rain occurs, the surface soil is remoistened and additional water moves through the previously moistened soil to increase the depth of recharged soil.

5. **Evaporative type of soil water regime** ($h_c < 1$) may be encountered in steppe, semidesert, and desert zones in localities with high water table. Capillary rise dominates in water movement. Saline ground waters make saline soils.

6. **Irrigative type of soil water regime**, as the name implies, is created on irrigated plots. Soil moisture is often maintained within the range of field capacity (FC) and the moisture of capillary discontinuity. But in some areas, different types of soil water regime may take place at different periods of time.

Plant uptake of water reduces the water content of the soil and greatly decreases hydraulic conductivity. As soil dries, water movement to roots becomes slower. Plants wilt when water movement to roots is too slow to satisfy the plant's demand for water.

During a year, the major changes in soil water content include: recharge of dry soil, surplus water and leaching, and depletion of stored water. Water deficits occur when water loss from the soil by evapotranspiration exceeds the precipitation plus stored water for a period of time.

19.7. Soil Water Management

Speaking on soil water management on agricultural land, H.D. Foth (1990) identified three basic approaches to it: 1) conservation of natural precipitation in subhumid and arid regions, 2) removal of water from wetlands, and 3) addition of water to supplement the amount of natural precipitation (irrigation).

Water conservation in areas with large water deficits (within arid and subhumid climates) should make use of techniques designed to increase the amount of water that enters the soil and make efficient use of this water. Modifying the infiltration rates is very important in this respect. High infiltration rates not only increase the amount of water stored for plant use but also reduce surface runoff (and soil erosion) and flooding. All practices improving soil structure increase infiltration rates. Infiltration is decreased by overgrazing and deforestation. Plant and residue cover favors the increase of infiltration rate and reduce raindrop impact and soil erosion, reduce surface runoff and water losses of the soil.

Water movement into and through a soil pore increases exponentially as the radius increases. A single earthworm channel that is open at the soil surface may transmit water 100 times faster than the same volume of smaller pores. Mulchings favor the population of earthworms in the surface layer. Many farmers modify the soil surface conditions with contour tillage and terraces. These practices temporarily pond the water and slow down the rate of runoff. This results in a longer time for the water to infiltrate the soil. Where there is little precipitation, runoff water is collected for filling reservoirs.

Summer fallowing is another practice to conserve water. But SOM losses are greater with it. After wheat is harvested, the land is left unplanted. No crop is grown and weeds are controlled by shallow cultivation or herbicides to minimize the loss of water by transpiration. The risk of soil erosion becomes greater. So the fallowing system is not 100 percent efficient. Saline seep may also be due to fallowing. Control of a saline seep depends on reducing the likelihood that surplus water will occur. Management practices to control saline seeps include reduced fallowing frequency, use of fertilizers to increase plant growth and water use, and the production of deep rooted perennial grasses (alfalfa) to remove soil vapor losses of soil moisture should also be regulated by mulches and residue cover.

Soil drainage is practical in areas with excessive soil wetness. The major purpose of drainage in agriculture and forestry is lowering the water table to increase the plant rooting depth. Some of the largest increases in plant growth resulting from drainage occur in forests. Few management practices are as effective for increasing trees growth.

Additional benefits of drainage in forests are easy logging, with less soil disturbance, and easy site preparation for the next crop (H.D. Foth, 1990).

Drainage of wet soils increases the length of the growing season in regions with low soil temperature and frost hazard. Seed germination is more rapid and roots grow faster.

Surface drainage is the collection and removal of water from the surface of the soil via open ditches. Surface drainage is used on low areas that receive a large amount of water from the surrounding higher land and where there are impermeable soils that have an insufficient hydraulic conductivity to dispose of excess water. Surface drainage ditches are used to dispose of excess water and to prevent saturation of the soil on the low end of irrigated fields. Subsurface drainage is achieved by drain lines which are installed in fields with trenching machines. Perforated plastic tubing is very popular. While draining some peat soils it is necessary to maintain their water table depth within 0.8-1.5 m.

Irrigation is an ancient agricultural practice. Today, approximately 11% of the world cropland is irrigated. Soils well suited for irrigation have an intermediate texture to a depth of 1 to 2 meters and without compact or water impermeable layers. Loamy soils that are underlain by coarse sands and gravel have greater retention of water in the overlying soil than soils that are loamy-textured throughout. The infiltration rate should be about 5 to 80 mm per hour and a moderate water storage capacity. The soil should not have an injurious soluble salt content. Irrigation water contains soluble salts and some provision (e.g., drainage) must be made to maintain a favorable salt balance to keep irrigated soils permanently productive. Salt accumulation results in formation of saline soils. High sodium saturation (greater than about 15% of the adsorbed cations) results in the formation of sodicity-affected or sodic soils, which can be improved by applying gypsum and then leaching the soil.

Questions and Assignments

1. Prove the importance of water for soil formation and productivity.
2. Characterize categories of soil water and its properties and comment on importance for crop growth.
3. What is soil water potential and what are its component parts?
4. What is the matric potential of soil water and how is it measured?
5. What is the soil-moisture characteristic curve and what practical applications of this curve do you know?
6. Speak on the factory determing the importance of osmotic potential.
7. What is a tensiometer and for which purposes can it be used?
8. Describe water movement in a saturated soil.
9. Describe water movement in an unsaturated soil.
10. What soil properties relate to water?
11. What is available water? How to compute and evaluate the amount of available water in the soil at field capacity (FC)?
12. Work out the general equation of water balance in the soil.
13. Name the types of soil water regime.
14. Name the basic approaches to soil water management in agriculture.

Chapter 20

Soil Air and Aeration

Soil aeration must be considered in relation to soil formation and plant growth. The process of soil aeration is one of most important determinants of soil productivity. Plant roots absorb oxygen and release carbon dioxide in the process of respiration. Adequate roots respiration requires that the soil itself be aerated. The gaseous exchange should take place between soil air and the atmosphere at such a rate as to prevent a deficiency of oxygen and an excess of carbon dioxide in the root zone (Fig. 93).

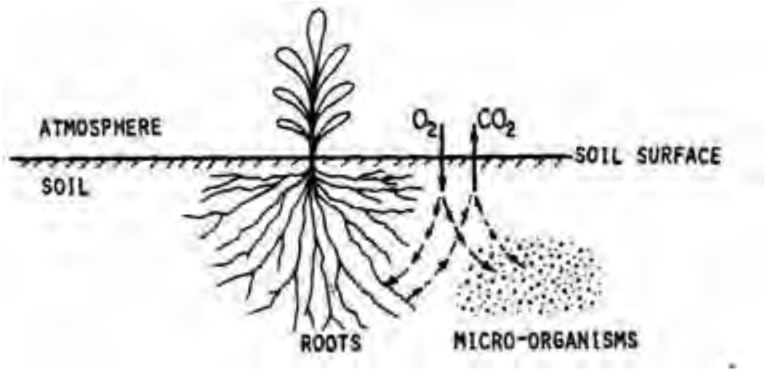


Fig 93. Soil aeration as a process of O_2 and CO_2 exchange with the atmosphere (schematic) (D. Hillel, 1982)

Soil microorganisms also respire and under conditions of restricted aeration, might compete with the roots of higher plants. Impeded aeration resulting from poor drainage and waterlogging, or from mechanical compaction of the soil, can strongly inhibit crop growth.

Soil air is the gaseous phase of the soil. It is the mixture of gases and vapors of organic compounds filling soil pores free from water. Soil air may be in the soil in three conditions (states): 1) free soil air, 2) adsorbed soil air, and 3) dissolved soil air.

But the needs of plants (plant roots) in oxygen are met mainly by free soil air as a result of aeration.

20.1. Composition of Soil Air

Soil air, despite its permanent interaction with the atmosphere, is characterized by a number of specific features. Atmospheric air composition is mostly unchangeable. In contains in percentage by volume: 78.08 of N_2, 20.95 of O_2, 0.93 of Ar, and 0.03 of CO_2. The composition of soil air is dynamic and inconstant. It may contain tens and hundreds times greater amount of CO_2 than atmospheric air. Oxygen concentration may reduce from 20.9 to 15-10% and lower.

In well aerated plow layers with favorable physical properties the content of CO_2 in soil air during plant growth does not exceed 1-2%, whereas the content of O_2 usually is not below 18%. In excessively wet heavy-textured cultivated soils the amount of CO_2 in soil air may reach 4-6% and more, whereas the amount of O_2 may fall below 17-15%. Swampy (peat) soils may have even higher CO_2 and lower O_2 concentrations in their air (Table 43).

Table 43. CO_2 (numerator) and O_2 (denominator) contents in Soil Air in June (% by volume)

Depth, cm	Soddy-podzolic soil, forest	Soddy-podzolic soil, spring wheat field	Soddy-podzolic gleyed soil, meadow	Chernozemic-meadow soil, flood plain	Meadow Gleyed Soil, meadow
	Grechin and Stratonovich, 1968			Ilyin and Shabanov, 1972	Grechin, 1964
10	$\frac{0.4}{20.5}$	$\frac{0.5}{20.4}$	$\frac{2.5}{17.7}$	–	$\frac{2.5}{18.3}$
25	$\frac{0.5}{20.6}$	$\frac{0.7}{20.3}$	$\frac{2.7}{17.1}$	$\frac{2.5}{18.5}$	$\frac{16.8}{2.9}$
50	$\frac{0.6}{20.2}$	$\frac{1.1}{19.9}$	$\frac{2.2}{17.5}$	$\frac{3.2}{17.8}$	$\frac{19.5}{2.2}$
100	$\frac{2.5}{18.3}$	$\frac{1.2}{19.9}$	$\frac{1.4}{19.2}$	$\frac{6.6}{13.8}$	–
150	$\frac{3.1}{17.5}$	$\frac{1.3}{18.7}$	$\frac{1.4}{18.8}$	$\frac{6.8}{13.4}$	–
200	$\frac{3.5}{17.5}$	$\frac{1.4}{19.5}$	$\frac{1.4}{19.0}$	$\frac{5.2}{15.5}$	–

Anaerobic conditions in the soil induce a series of reduction reactions, both chemical and biochemical. Included among these reactions are denitrification ($NO_3^- - NO_2^- - N_2O - N_2$), manganese and iron reduction, and sulfate reduction to form hydrogen sulfide. Some of the numerous products are toxic to plants (ferrous sulfide, ethylene, acetone, butyric, and phenolic acids).

Microbial decomposition of organic residues accounts for the major portion of the CO_2 evolved. Incorporation of large quantities of manure, especially when moisture and temperature are optimum, will alter the soil air composition appreciably.

There is a marked seasonal variation in the composition of soil air. This variation can be accounted for by soil moisture and soil temperature differences. High moisture tends to favor a low oxygen and a high carbon dyoxide levels in the soil air. In temperature regimes this situation often prevails in winter and late spring when soil moisture is generally higher.

Because soils are normally drier during summer months, conditions for gaseous exchange are better during this period. Some exceptions to this rule may be found, however, since high summer temperatures also encourage rapid microbial release of CO_2.

One more another interesting difference between the atmosphere and soil air is that the latter is characterized by a high relative humidity (nearly 100%), except at the soil surface during prolonged dry spells.

20.2. Gas Exchange of Soil Air with Atmosphere

The exchange of air between the soil and the atmosphere can occur by means of two different mechanisms: **convection** and **diffusion.** By the linear rate law, the flux is proportional to the moving force.

In the case of **convection** (mass flow), the moving force consists of a gradient of total gas pressure. It results in **the entire mass of air streaming from a zone of higher pressure to one of lower pressure.**

A number of phenomena can cause pressure difference between soil air and the external atmosphere. Among these phenomena are: 1) barometric pressure changes in the atmosphere, 2) temperature gradient, and 3) wind gusts over the soil surface. Water infiltration can cause displacement of soil air, the fluctuations of a shallow water table can push it upwards or draw it downwards. The rapid extraction of soil water by plant roots also draws the atmospheric air into the soil. Short-time changes in soil air pressure can also occur during tillage or compaction by machinery. It is possible to write down an equation for the convective flow of air in the soil as an one analogous to Darcy's law for water flow, as follows:

$$g_V = -(K/\varsigma)\nabla P \tag{57}$$

where g_V is the volume of convective flux of air (through a unit cross-sectional area per unit time),

K is permeability of the air- filled pores,

ς is viscosity of soil air, and

∇P is the three-dimensional gradient of soil air pressure. In one dimension, this equation takes the form:

$$g_V = -\left(\frac{K}{\varsigma}\right)\frac{dP}{dX} \tag{58}$$

In the case of diffusion, the moving force is a **gradient of partial pressure** (or concentration) of any constituent member of the variable gas mixture which is called *air*. It causes the molecules of the unevenly distributed constituent to migrate from a zone of higher to lower concentration even while the gas as a whole may remain isobaric and stationary.

The diffusive transport of gases such as O_2 and CO_2 in the soil occurs partly in the gaseous phase and partly in the liquid phase. Diffusion through air-filled pores maintains the exchange of gases between the atmosphere and the soil. Diffusion through water films of various thicknesses maintains the supply of oxygen to, and disposal of carbon dioxide from, live tissues, which are typically hydrated. For both portions of the pathway, the diffusion process can be described by Fick's law:

$$g_d = -Dd_C/d_X \tag{59}$$

where in g_d is the diffusive flux (mass diffusing across a unit area per unit time),

D is the diffusion coefficient (the dimension of areas per time);

C is concentration (mass of diffusing substance per volume);

X is distance; and

d_C / d_X is the concentration gradient.

D. Hillel in his “Introduction to Soil Physics” notes that in the air phase of the soil the diffusion coefficient, D_S must be smaller than that in bulk air D_0, owing to the limited fraction of the total volume occupied by continuous air-filled pores and also to the tortuous nature of these pores. D_S is the function of the air-filled porosity (f_a). J. Penman (1940) found a linear relation:

$$D_S / D_0 = 0.66 f_a \tag{60}$$

wherein 0.66 is a tortuosity coefficient. J. Penman’s equation proved useful for practical computations and became very popular, though there are a numerous of other equations. In aggregated soils, gaseous diffusion takes place rapidly in the interaggregate macropores. Such pores quickly drain out after a rain or irrigation and form a network of continuous air-filled voids.

The condition of air exchange depends very much on the diffusion coefficient (Table 44), as well as on **the air-related soil characteristics.** These characteristics used in Ukraine, but not mentioned elsewhere include air penetrability and air capacity of the soil.

Table 44. CO_2 Diffusion Coefficients at Atmospheric Pressure (101 KPa) as Determined by N.P. Poyasov

Soil	ºC	Air-filled porosity, % of total volume	D_0	D_S	D_S / D_0
Ordinary chernozem	21	61.6	0.161	0.0473	0.294
Ordinary chernozem	21	37.3	0.160	0.0239	0.149
Ordinary chernozem	21	21.7	0.160	0.0160	0.100
Soddy-Podzolic	18	26.9	0.152	0.0123	0.078
Soddy-Podzolic	18	26.7	0.159	0.0106	0.067
Soddy-Podzolic	18	6.7	0.158	0.0009	0.006

Air penetrability is the ability of the soil to be penetrable to the air. It is measured by the amount of the air (ml) that penetrated under a definite pressure through the unit cross-sectional area (1 cm^2) and a unit of soil thickness (1 cm) per unit of time (min). Air penetrability depends on soil texture, bulk density, wetness and aggregation. Soil air moves through the pores which are not filled with water and not isolated from each other. The larger the pores, the better the air penetrability.

Air capacity characterizes the content of air in the soil expressed in percentage of the total soil volume. The amount of the air depends on soil wetness and its total porosity. The greater the porosity and the lower the wetness, the greater the air capacity of the soil. As soils in the field are always wet, their air capacity is a dynamic characteristic. The air-filled porosity at field capacity of soil moisture (FC) in percentage of total soil volume should not be less than 15 for mineral and 30-40 for organic (peat) soils.

These figures determine the aeration requirement for a given soil. But it is important to acquire quantitative knowledge of the aeration requirements for different crops and soils in varying circumstances. Alexander (1961) notes that such information is difficult to obtain, as the rate and spatial distribution of soil respiration, as well as its temporal variation, depend on numerous factors, among which are temperature, soil wetness, organic matter, and the time – variable respiratory activity of both macro- and microorganisms.

Anaerobiosis, or oxygen stress, will occur in the soil whenever the rate of supply falls below the demand. This condition can develop very quickly, since the storage of oxygen in the soil is generally rather low in relation to the quantity required for soil respiration.

20.3. Soil Air Regime and its Regulation

Soil air regime is defined as the set of all phenomena of atmospheric air entering the soil, its movement within the soil profile, changes in its composition and physical condition during the interaction of soil air with solid, liquid, and live phases of soil as well as the exchange of air between the soil and the atmosphere.

Soil air regime is subjected to diurnal, seasonal, and annual dynamics and is directly influenced by the physical, chemical, physico-chemical and biological soil characteristics. It depends upon climate and weather, character of vegetation, cultivated crops, and the practices (technologies) of crop production systems.

Seasonal differences in the composition of soil air and the character of gas exchange can be accounted for by soil – moisture and soil – temperature variation. All aerobic organisms are unable to function properly in the absence of free oxygen. The rate of soil OM and residue oxidation is slow (in swampy areas). Symbiotic N-fixers and some of the nonsymbiotic (Azotobacter) groups fail to function properly. Only the anaerobic and facultative organisms function properly under poor aeration conditions. They are able to do so because of their ability to utilize combined oxygen. They produce reduced forms of certain elements such as Fe and Mn, which are often toxic to higher plants. Organic toxins may also develop. The conditions of poor aeration deeply affect the growth of plants, particularly the roots, the absorption of nutrients, the absorption of water, and favor the formation of certain inorganic and organic substances, toxic to plant growth.

Some crops, for example, tomatoes, require high air porosities (up to 20-30%) for best growth. Timothy grass, if well supplied with nitrogen, can grow with very low air porosity, and rice grows normally submerged in water. Sugar beets are also very sensitive to inadequate aeration, alfalfa and legumes.

Anaerobic decomposition is much slower in poor aeration and its products are different from aerobic:

$$C_6H_{12}O_6 \rightarrow 3CO_2 + \underset{methane}{3CH_4}$$

Less complete decomposition provides gives other products such as organic acids, which, under extreme conditions, may accumulate in toxic quantities. The nitrogen compounds remain in reduced forms. They are not subject to nitrification.

In general, it can be stated that oxidized forms are much more desirable for most plants. This is especially true if the poor aeration occurs in an acid soil. Lack of lime intensifies the adverse affects of poor aeration or iron and manganese toxicities, and the soils become gleyed quicker. Poorly drained soils usually show a condition wherein alternate streaks and mottles of oxidized and reduced materials occur. This “mottled” condition is not conducive to proper plant growth.

The improvement of aeration is badly needed by many soils but especially by those with temporary or permanent excessive wetness.

Soil air regime is regulated with the aid of crop production technologies as well as by soil reclamation special practices.

All practices improving or conserving soil structure will favor the regime of soil air and aeration, including minimum tillage, stubble mulch cultivation, farming on the contours, employment of perennial grasses in crops rotation, chemical amendment, sufficient manuring, green manuring, special mole-tunnel tillage, and, finally, the use of synthetic soil granulators. Soil crust must not be allowed to form, or, if formed, must be quickly broken.

Artificial drainage is appropriate in some places but inappropriate elsewhere. Some wetlands must be protected and not used as cropland, and therefore not subject to drainage. Some drained lands become so alkaline from sodic conditions or so acid from the oxidation of sulfides that nothing will grow on them. Artificial drainage should be provided only where soil conditions are suitable, and the land use will benefit.

Questions & Assignments

1. What are the peculiarities of soil air compared to atmospheric one?
2. What is the function of soil air and aeration for plants growth and soil formation?
3. What is the gas exchange in the soil and though which mechanisms does it occur?
4. Name and characterize the aeration-related properties of the soil.
5. How can soil aeration optimal conditions be described?
6. What is soil air regime and how can it be regulated?
7. Name and characterize soil aeration improvement practices involved in crop production technologies.
8. Characterize artificial drainage as a means of improving soil aeration.

Chapter 21

Soil Temperature and Heat Flow

Soil temperature determines the rates and directions of soil physical processes, the energy and mass exchange of the soil with the atmosphere, including evaporation and aeration. It also strongly influences biological processes (seed germination, seedlings emergence and growth, roots development, and microbial activity.

Soil temperature depends on changes in the radiant, thermal, and latent **energy exchange processes.** The effect of these **exchange processes** taking place through the soil surface are propagated into soil profile by a series of **transport processes.** The understanding of the soil heat, or rather thermal regime, requires a through knowledge of the processes at play and of the environmental and soil parameters which govern their rates. The essential soil properties related to heat flow and the formation of thermal regime include the specific and volume heat capacity, thermal conductivity, and thermal diffusivity. These soil parameters are deeply affected by bulk density, soil granulation, wetness and texture. Solar radiation is by far the most important source of heat in the soil. To understand how it affects soil temperature, it is necessary to consider the modes of energy transfer from basic physics.

21.1. Modes of Energy Transfer

There are three principal modes of energy transfer: 1) radiation, 2) convection, and 3) conduction.

The emission of energy in the form of electromagnetic waves from all bodies above 0°K is called **radiation.** The total energy emitted by a body is proportional to the fourth power of the body's surface absolute temperature (Stephan-Boltzmann law).

The absolute temperature also determines the wave length distribution of the emitted energy. The wave length of maximal radiation intensity λ_m is inversely proportional to the absolute temperature (Wien's law).

$$\lambda_m = \frac{2900}{T} \tag{61}$$

where λ_m is expressed in microns.

The radiation emitted by the soil surface has its peek intensity at a wave length of about 10 μ_m and its wave length distribution over the range of 3-50 μ_m. This is in the realm of **infrared or heat radiation.** The sun's surface radiation at the effective temperature of about 6000°K includes the visible light range of 0.3-0.7 μ_m, as well as some infrared radiation of greater wavelength (up to about 3 μ_m), and some ultraviolet radiation ($\lambda\langle 0.3\mu_m$). That is why it is customary to name the incoming solar spectrum *short-wave radiation*, and the spectrum emitted by the earth *long-wave radiation*.

The second mode of energy transfer, called *convection*, involves the movement of a heat-carrying mass, like an infiltration of warm water into an initially cold soil. The third mode of energy transfer, called *conduction*, is the propagation of heat within a body by internal molecular movement. Temperature is an expression of the kinetic energy of a body's molecules. The existence of temperature differences within a body will normally cause the transfer of kinetic energy by the numerous collisions of rapidly moving molecules, from the warmer region of the body to its colder region.

The process of heat conduction is analogous to the process of diffusion: heat conduction tends to equilibrate a body's internal distribution of molecular kinetic energy (to equilibrate its temperature i.e.).

There is still another mode, that of latent energy transfer, say the heat-releasing stage of water vapor condensation or transition back and forth from ice to liquid water.

Within the soil, heat transfers by radiation, convection, and distillation are generally of secondary importance. The primary process of heat transport is by molecular conduction.

The first law of heat conduction, known as Fourier's law, states that the flux of heat in a homogeneous body is in the direction and proportional to the temperature gradient:

$$g_h = -K\nabla T \qquad (62)$$

where g_h is the thermal flux (i.e., the amount of heat conducted across a unit cross-sectional area in unit time), K is thermal conductivity, and ∇T is the spatial gradient of temperature**.** In one- dimensional form, this law is written as follows:

$$g_h = -K_x \frac{d_T}{d_x} \qquad (63)$$

where d_T / d_x is the temperature gradient in any arbitrary direction designated x. The negative sign in these equations is due to the fact that heat flows from a higher to a lower temperature.

If g_h is expressed in calories per square centimeter per second and the temperature gradient in degrees Kelvin per centimeter, K has the units of calories per centimeter-degree second (Cal/cm °K sec).

21.2. Heat – Related Properties of the Soil

In Ukrainian soil science it is customary to identify the following heat-related properties (characteristics) of the soil: 1) heat absorbing capacity of the soil; 2) heat capacity of the soil (specific, volumetric, and effective (specific or volumetric)); and 3) **heat** conductivity of the soil with an ever increasing tendency to substitute it by the **thermal** conductivity, which is more agronomically important.

Heat absorbing capacity is the ability of the soil (and land) surface to absorb (or reflect) the incoming short-wave solar radiation. This capacity is evaluated by the Albedo characteristics, expressed in the relative amount (percentage) of reflected short-wave solar radiation of its total amount coming to the surface. Albedo (A) is determined by the ratio:

A = solar short-wave radiation reflected from the surface × 100 / total amount of short-wave solar radiation A.P. Chudnovskiy (1959) gives the following examples of Albedo for the various surfaces (Table 45).

Table 45. Examples of Albedo for various surfaces

Object of measurement	A	Object of measurement	A
Chernozem soil (dry)	14	Spring wheat field	10–25
Chernozem soil (moist)	8	Winter wheat field	16–23
Grey desert soil (dry)	25–30	Green grasses	26
Grey desert soil (moist)	10–12	Dried grasses	19
Clay (dry)	23	Cotton field	20–22
Clay (wet)	16	Rice	12
Sand white and yellow	34–40	Potato field	19

Soils rich in OM absorb more solar radiation than light colored soils; wet soil surface absorbs more solar radiation than dry.

Heat capacity (C) is the ability of the soil mass to absorb heat. **The volumetric heat capacity (C_v)** of a soil is defined as the change in heat content of **a unit bulk volume** of soil per unit change in temperature. Its units are calories (or joules) per cubic centimeter per degree, or joules per cubic meter per degree (Kelvin).

Specific heat capacity (Cm) of a soil is defined as the change in heat content of a unit mass of soil per unit change in temperature. It is measured in the same units as volumetric heat capacity.

Heat capacity depends on mineralogical and textural composition of the soil, OM content, wetness, porosity, and aeration porosity (Table 46).

Table 46. Heat Capacity of Soil Components and Some Minerals

Substance	Heat capacity	
	Specific, cal /gm. °K	Volumetric, cal /cm^3. °K
Quartz sand	0.196	0.517
Clay	0.233	0.577
Peat	0.477	0.611
Water	1.000	1.000
Quartz	0.198	–
Kaolin	0.233	–

As the heat capacity of water is higher compared to the mineral and organic components of the soil, it requires more heat to increase the temperature of wet soils as compared to dry. Wet soils get warmer more slowly and give out their heat more slowly whereas dry soils get warmer and cooler more quickly.

The value of C can be calculated by addition of heat capacities of various constituents, weighed according to their volume or mass (weight) fractions. In typical mineral soils, the volume fraction of solids is in the rage of 0.45-0.65 and C_V values range from less than 0.75 cal/cm3 °K in a dry state to about 0.25 cal/cm3 °K in a water – saturated state.

Volumetric heat capacity may be calculated by multiplying the respective specific heat capacity by the particle density of a mineral or any other constituent of the soil.

The effective heat capacity is that of a wet soil in the field. The specific effective heat capacity is very often calculated in Ukraine according to P.I. Andrianov’s formula:

$$C_m ef = 0.2x + 0.7y + (W - y) \tag{64}$$

where $C_m ef$ is in cal/gm; x is the mineral part content in the soil in parts by weight; y is the content of matrix-bound (adhesion) water in the soil in the same units, and W is soil wetness in the same units.

Clay soils in a wet state get warmer more slowly in spring than sandy soils. Sandy soils accumulate less heat and give it out much quicker than clay soils in cold weather. Well structured soils rich in OM are cooler than poorly aggregated ones.

Heat conductivity is the soil’s ability to transport heat. Specific heat conductivity, designated K, is defined as the amount of heat transferred through a unit area in unit time at a unit temperature gradient. Heat conductivity determines the rate with which heat is transported from one (warmer) soil layer to another (cooler). Soil components are characterized by different values of heat conductively (cal /see cm. °K):

Air	0.00006
Water	0.00136
Peat	0.00027
Quartz	0.00240
Granite	0.00820
Basalt	0.00520

Heat conductivity of the mineral part of the soil is 100 times higher than that of the air. Heat conductivity of water is only 28 times higher than that of the air. The wetter the soil, the higher its heat (and thermal) conductivity, the looser (less compact) the soil, the lower its heat conductivity. In summer, when the surface layer of the soil becomes dry, its thermal conductivity rapidly decreases reducing thereby heat transfer from the surface layer downward. Water accumulated in soil in full favors the storage of heat in soil which protects the seedlings of winter crops from early frosts.

21.3. Heat (Thermal) Regime of Soil

Thermal regime of the soil is formed under the influence of atmospheric climate (the flux of solar radiation, humidity, continentality, etc.) as well as under the influence of topography (land forms), vegetation and snow cover (*Soil Science* edited by I.S. Kaurichev). Soil temperature is the main soil characteristics evaluating its “heat condition”.

In nature, soil temperature varies continuously in response to the ever-changing meteorological regimes acting upon the soil – atmosphere interface. That regimes are characterized by a regular periodic succession of days and nights, and of summers and winters. Yet the regular diurnal and annual cycles are perturbed by such irregular episodic phenomena as cloudiness, cold waves, warm waves, rain storms or snowstorms, and periods of drought. The thermal regime of soil profiles can be complex indeed, yet not altogether unpredictable” (D. Hillel, 1982).

Thermal regime in concomitance with air and water relations effects soil formation as well as soil fertility and crop productivity. The seeds of various crops germinate only within a certain range of soil temperatures (Table 47).

Table 47. Temperature Ranges for Seed Germination (ºC)

Crop	Minimum	Optimum	Maximum
Wheat, barley, oat, and rye	0–5	25–31	31–37
Sunflower	5–10	31–37	37–39
Cotton, rice, and pumpkins	16–24	37–44	44–50

Soil temperature affects the activity of soil microorganisms. For most of them optimum temperature is within 25-30 ºC. Solar radiation energy is an essential source of heat in the soil. It is absorbed by the soil surface and transformed into heat energy. Green plants accumulate solar energy in the form of organic matter.

Any soil type is characterized by a definite dynamics of temperatures at different depths in the profile during the period of crop growth. The widest ranges of temperature fluctuations are characteristic for the surface soil layers. With increasing depth, the fluctuation becomes less pronounced. Diurnal changes in soil temperature completely subside at the depth of 40-50 cm.

Annual soil temperature dynamics depends on the lattitude of a locality (or soil-geographic zone). On chernozemic soils in winter months at the depth of 30-40 cm the temperature drops below 0 ºC, in June-August it reaches a maximum value, and then decreases again towards winter.

At greater depths the annual temperature fluctuations become insignificant. The depth of soil freezing in winter depends on the thickness of a snow cover. Below the snow layer, the soil becomes frozen only to a small depth, whereas in snowless winters the depth of soil freezing may reach 0.7-0.9 meters and more. Snow-retention may thus lead to soil moisture accumulation and heat conservation. While choosing land areas for agricultural fields, account should be taken of soil properties, including texture, topography, and hydrothermal conditions of a locality.

Fig. 94.Temperature Measurement in typical Chernozem (E. Berezhniak, 2014)

Heat balance of a given plot or a locality, as the quantitative expression of their thermal regimes, is to a very great extent determined by the radiation balance. Actually, the thermal (or heat-related, as it is customarily called in Ukrainian soil science and agronomy) regime of the soil is understood as a set of phenomena of the supply, transport, and expenditure of heat by the soil. Solar radiation received at the soil surface is partly reflected back into the atmosphere (albedo) and partly absorbed by the soil surface. As was already stated, a dark-colored soil and light-colored quartz sand may absorb about 80 to 30 % of the incoming solar radiation, respectively. Heat is lost from the soil due to water evaporation, radiation reflected back into the atmosphere, heating of the air above the soil, and heating of the soil. For the most part the gains and losses are in balance. But during the daytime or in summer the gains exceed the losses. The reverse is true for the nights and winters.

Heat balance is composed of the radiation balance (T_b), turbulent flux of heat caused by heat exchange between the soil surface and the air (T_k), heat expended for the evapotranspiration ($T\gamma$), and heat exchange between the soil layers (T_n). Heat balance equation admits the algebraic equality of various fluxes:

$$T_b + T_k + T\gamma + T_n = 0 \qquad (65)$$

Considering mean annual soil temperatures and the character of soil freezing, V.N. Dimo (1972) identified 4 types of soil thermal regimes: permafrostic, prolonged seasonal freezing, seasonal freezing, and non-freezing.

There is no **permafrostic thermal regime** on the territory of Ukraine. It is characteristic of the polar Eurasian regions and of eastern Siberia. Mean annual temperature in the soil profile is negative. Soil freezing occurs to the depth of permafrost. Prolonged seasonal freezing is characterized by soil freezing to the depth of no less than 1 meter, but the soil freezing depth does not reach the permafrost layer. **Seasonal freezing** is characterized by positive (above freezing point) annual meanings exceeding 5 months. This type of soil thermal regime is most typical for the territory of Ukraine.

Nonfreezing type of soil thermal regime is found in the southern regions where the soils do not freeze at all. It is important to remember that the maximum temperature of a dry surface soil may definitely exceed that of the air, but in winter even surface soils do not fall greatly below freezing.

Soil thermal regime regulation. Various practices in crop production systems, soil (and land) reclamation, and agrometeorological procedures can be employed for the soil thermal regime regulation. Some practices of soil tillage may include soil ridging, compacting, stable mulch cultivation, snow-trapping practices, etc.

Soil reclamation practices include irrigation, drainage, plantings of windbreaks and shelterbelts, and the practices of drought control. Agrometeorology may be constructive with the practices of radiation and insulation control, smokescreens, etc. Cultivation on the ridges improves soil warming, intensifies heat exchange between the air and the soil, and increases the crop resistance to frost. Soil compacting may increase the mean daily temperatures in the layer 10 cm thick, which directly underlies the compacted layer of soil, by some 3-5 ºC. Mulching reduces the reflection of radiation from the soil surface. Black mulches reduce soil albedo by 10-15%, whereas white materials are employed to reduce the excessive heating.

Shelterbelts improve snow trapping and accumulation, and snow warms up the soil, reducing the vertical heat exchange between the near-surface and the atmospheric air. This reduces the air temperatures between the belts in the day and increases them during the night hours. Irrigation reduces the radiation reflection by 20% which increases the influx of heat energy to the soil. Organic manures and composts also increase soil temperature. Rows of high-standing plants (maize, sunflower, etc.) create a “hot-house effect” accompanied by the increases in soil temperature. This practice is employed in the heat-deficient regions to increase the yields of vegetables.

Questions & Assignments

1. Speak on the role of soil temperature in soil formation and the productivity of crops.
2. What modes of energy transfer do you know? Which of them is the most important for soils?
3. What heat related properties of the soil do you know?
4. Characterize albedo for the various surfaces. How can it be employed for the soil temperature regulation?
5. What is the difference between the specific and volumetric heat capacities of the soil? How can one be calculated from the other?
6. What is soil heat conductivity? What are its values for the various components of the soil?
7. Define heat (thermal) regime of soils. In what way can it be expressed quantitatively?
8. Speak on temperature ranges for seeds germination.
9. Why is heat balance very much dependent on radiation balance?
10. Speak on the practices of soil thermal regime regulation.

Chapter 22

Soil Solution and the Redox Processes in Soils

Rainwater entering the soil contains certain amounts of dissolved substances: atmospheric air gases (O_2, CO_2, N_2, NH_3, etc.) and the components of the solid particulates in the air represented mainly by the dust particles. In the soil this water interacts with solid phase so that some soluble components dissolve in it. The water in the soil is thus not pure water but the **soil solution.**

Soil solution is of a very great importance in soil genesis and development of productivity. The actively participates in processes of mineral transformation and those involving organic residues and all other forms of SOM. Due to it a continuous transport of substances and energy takes place in the soil profile. Soil solution is exceptionally important for plant nutrition. It is therefore necessary to know its composition and properties (reaction, buffering capacity, osmotic pressure, etc.), as well as their dynamics.

22.1. Methods of Soil Solution Separation

Diverse procedures are employed for the soil solution separation from the solid phase of the soil – pressing out, centrifuging, and replacement by the other liquids. The amount of soil solution separated depends of the ability of the soil to retain water and the extent of soil wetness. The separation of soil solution by centrifuging is possible only with the soils saturated to nearly their full capacity.

The extraction of soil solution by replacing it with an apolar liquid is performed by pouring this liquid over the soil sample with natural wetness in a special extraction column. Ethyl alcohol is often used for the purpose.

Lysimeters may also be employed for soil solution separation or, more exactly, for the collection of liquid infiltrated through the soil under natural (in situ) or laboratory conditions. The procedure is convenient for studying the solutions of soils such a podzolic, soddy podzolic, or oxisolic ones where saturated flow occurs.

Soil solution can be investigated in the field using portable ionometers with ion-selective electrodes. The use of soil pastes (at soil wetness equal to full capacity) or soil water extracts is the easiest but the least precise procedure to study soil solution. But such a procedure is still employed to study saline soils.

Soil solution reaction (pH), pNa, its electrical conductivity, and the redox potential should be measured directly in the field.

22.2. Composition and Concentration of Soil Solution

H.D. Foth in the *Glossary of his Fundamentals of Soil Science* defines soil solution as the agneous liquid phase of the soil and its solutes. Soil solution is in constant interaction with solid and gaseous phases of the soil as well as with plant roots, so its composition and concentration are the results of biological, physico-chemical, and physical processes underlying this interaction. The rate and direction of soil processes are subject to a considerable seasonal variability, so the composition of soil solution is very dynamic.

The concentration of soil solution is usually not high, not exceeding a few grams per liter. Saline soils may be an exception. Their solutions may contain tens and even hundreds of grams per liter.

Soil solution may contain mineral, organic, and the intermediate type, organic-mineral substances represented by ionic, molecular, and colloidal forms. Soil solution also contains dissolved gases like CO_2, O_2, etc.

Of inorganic compounds soil solution may contain anions like HCO_3^-, CO_3^{2-}, NO_3^-, NO_2^-, SO_4^{2-}, Cl^-, $H_2PO_4^{2-}$, and cations like Ca^{2+}, Mg^{2+}, Na^+, NH_4^+, K^+, H^+, Al^{3+}, Fe^{3+}, Fe^{2+}, etc (Table , meq per 100 g soil / %).

Table 48. Soil Extract Concentration of Peat Soil of Trubizh Floodplane (E. Berezhniak, J. Zeitz, 2005)

Type of soil	Depth (cm)	Anions				Cations					
		HCO_3^-	Cl^-	SO_4^{2-}	Sum of anions	Ca^{2+}	Mg^{2+}	Sum Ca^{2+} + Mg^{2+}	K^+ + Na^+	Sum of cations	Sum of harmful salts
Peat -clay soil	0-10	2,69 0,164	2,80 0,099	0,037 0,0017	5,527 0.265	2,24 0,045	1,792 0,034	4,032 0,079	1,495 0.034	5,527 0,113	0,195
	10-20	3,36 0,205	3,70 0,131	0,037 0,0017	7,10 0.338	1,344 0,027	0,448 0,005	1,792 0,032	5,308 0.122	7,100 0,154	0,383
	30-40	4,46 0,273	2,24 0,080	0,052 0,0025	6,75 0.3555	0,896 0,018	0,448 0,005	1,344 0,023	5,406 0.124	6,750 0,147	0,269

Organic compounds in soil solution are represented by the soluble components of plant residues, products of their decomposition, as well as the metabolism products of higher plants, microorganisms, and soil fauna, soluble humic substances, carbohydrates, alcohols, amino acids, enzymes, tannins, etc. Organic-mineral derivatives are represented mainly by the complex compounds of humus acids with polyvalent cations, like fulvates of aluminum, iron, and their hydroxy cations.

Osmotic pressure of soil solution is of primary importance to plant roots. If osmotic pressure of soil solution exceeds that of root cells the roots are unable to absorb water from the soil. Osmotic pressure of soil solution depends on its concentration and the extent of dissociation of the molecules dissolved in it. The highest values of osmotic pressure are found in the soil solution of very saline soils, especially those with heavy texture and pronounced adsorbing capacity.

Different soil types, as well as the different horizons of the same soil profile are characterized by different values of osmotic pressure of their solution (Table 48).

Table 48. Osmotic Pressure of Soil Solution (A.G. Gebgard)

Soil	Horizon	Osmotic Pressure of Soil Solution, Pa	Soil	Horizon	Osmotic Pressure of Soil Solution, Pa
Chernozem	$A+B_1$	2.05×10^5	Wet solonchak	First	11.2×10^5
(Mollisol)	B_1	1.68×10^5	(very saline soil)	Second	11.6×10^5
	B_2	3.80×10^5	Solod	A_2	0.20×10^5
Solonetz	A	1.56×10^5	(solot)	B_2	1.90×10^5
“Columnar”	B_1	0.40×10^5		B/C	1.56×10^5
(Sodic soil)	B_2	0.38×10^5			

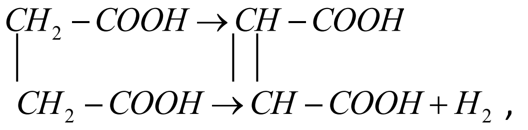
There is some research evidence that increased values of osmotic pressure in the soil reduces the period of crop growth, detains tillering, but accelerates flowering and spike formation in wheat. The grain reaches maturity more quickly and contains more protein, remaining undersized. Higher osmotic pressure reduces transpiration coefficient (the expenditure of water per a unit weight of dry matter). Osmotic pressure is measured using cryoscopic procedure by the freezing point of soil solution.

22.3. Oxidation and Reduction Processes in Soil Solution

H.L. Bohn, et al in their *Soil Chemistry* state that Redox reactions in the soil are the result of a cycle started by photosynthesis. The soil reactions complete the cycle as they utilize the energy stored by photosynthesis, dispose of organic wastes, and produce CO_2 needed for additional photosynthesis. Such point of view is rather interesting and new. Modern Ukrainian textbooks on soil science still regard this matter in a more traditional way. Anyway, soil should be considered as a complicated oxidation-reduction system. Oxidation processes are those of oxygen addition:

$$2KNO_2 + O_2 \rightarrow 2KNO_3,$$

hydrogen subtraction:



$$\begin{array}{lcl} CH_2-COOH & \rightarrow & CH-COOH \\ | & & \| \\ CH_2-COOH & \rightarrow & CH-COOH + H_2 \end{array},$$

and the processes of electron subtraction without any participation of hydrogen or oxygen:

$$Fe^{2+} - e = Fe^{3+}.$$

The processes opposite to these are known under the name of reduction processes. In the general scheme of redox processes, oxidation is the losing and reduction – the gaining of electrons:

$$Ox + e \rightarrow \mathrm{Re}d; \quad \mathrm{Re}d - e \rightarrow Ox.$$

Oxidation and reduction processes are very important for the transformation of soil organic matter. Humification is generally an oxidation process. The majority of oxidation reactions involving the organic substances of the soil are irreversible. Oxygen is the electron donor, and carbon is the electron acceptor.

Redox reactions involving carbon, nitrogen, and sulfur are determined largely by the availability of electrons and are usually catalyzed by enzymes.

But the reversible oxidation-reduction reactions are represented in soils by such processes as oxidation and reduction of iron $(Fe^{3+} \Leftrightarrow Fe^{2+})$, manganese $(Mn^{4+} \Leftrightarrow Mn^{2+})$, nitrogen $(N^{5+} \Leftrightarrow N^{3+})$. Oxidation and reduction of oxygen $(O_2 \Leftrightarrow 2O^{2-})$, hydrogen $(H_2 \Leftrightarrow 2H^{+})$, and sulfur $(S^{6+} \Leftrightarrow S^{2-})$ are also found in the soil environmental. The majority of those reactions have a biochemical nature and are closely linked with the microbial processes. The major electron donors in soils are freshly fallen plant matter and soil organic matter (SOM). SOM is enriched in carbon with respect to plant matter. The carbon groups that form in soil humus tend to be more aromatic, and less rich in oxygen, than plant matter. Other electron donors in soils besides organic carbon include the nitrogen and sulfur in amino $(-NH_2)$ and sulfhydral $(-SH)$ groups, and the ammonium ions in organic matter. Soil microorganisms create other electron when oxygen is lacking in the soil.

The role of soil in oxidation-reduction reactions is to provide electron acceptors for the oxidation of organic compounds. Oxygen is the strongest electron acceptor for this process. A high oxygen demand is normally caused by the presence of readily decomposable organic compounds or by large numbers of microorganisms and their considerable activity. The oxygen demand of soil can exhaust the dissolved oxygen in a waterlogged soil in less than 24 hours. The oxygen of soil air and soil solution is the major electron acceptor in the redox processes but the role of dissolved oxygen is very little. Soil aeration becomes insufficient with increase in soil wetness, soil compaction, and crust formation. The lowest redox potential is found in soils with wetness close to full saturation. According to I.P. Grechin, at optimum temperature and soil moisture the transition from aerobic to anaerobic conditions in the plowlayer of a soddy-podzolic soil occurred with the content of oxygen in soil air within 2.5-5%.

In the experiments conducted by M.V. Kurlykowa, under perennial grasses in a soddy-podzolic soil the stable anaerobiosis occurred with soil wetness over 80% of full capacity and aeration porosity of 6% by volume. Increases in SOM content favored a more rapid reduction of oxidation-reduction potential in excessively wet soils compared to their variants with lower OM content.

The property of substance to donate or accept electrons is given by its electrode potential. High electrode potential means that the elements or ions would readily accept electrons. Low electrode potentials mean that the elements or ions donate electrons instead.

Redox potential (E) is used for the quantitative evaluation of the oxidation-reduction condition in the soil. It reflects the total effect of various oxidation-reduction systems in the soil at a given moment of time. It is characterized by the following equation:

$$E = E_0 + \frac{RT}{nF} \ln \frac{(Ox)}{(\mathrm{Re}d)} \tag{66}$$

where E_0 is the standard electrode potential, R is the gas constant, T is the absolute temperature, and F is the Faraday constant.

As is evident from the equation, the greater the concentration of the oxidizer (Ox), the higher the redox potential. The redox potential in relation to hydrogen electrode (Eh) is characterized by the formula:

$$Eh = E_0 + \frac{RT}{nF} \ln \frac{[H^+]^2}{[H_2]} \tag{67}$$

E_0 for hydrogen electrode is conventionally accepted to be equal to zero.

$n = 2$ $(H_2 - 2e = 2H^+)$, substituting the known values of R and F into the formula and taking T=291K, the above equation may be simplified:

$$Eh_{(v)} = 0.029 \log \frac{[H^+]^2}{[H_2]}. \tag{13.3}$$

To obtain Eh data comparable at different pH values, T.J. Clark proposed to employ the rH_2 index computed by the formula:

$$rH_2 = \frac{Eh}{29} + 2pH. \tag{68}$$

with $rH_2 > 27$, oxidation processes assuredly dominate over reduction ones, whereas with $rH_2 < 27$, the reduction processes prevail, which become particularly intensive with $rH_2 < 20$.

An alternative expression of the electrons potential uses the term pe. The relation between pe and Eh at 25°C is:

$$Eh = 0.059 pe \tag{69}$$

Under conditions reducing, substances such as Fe^{2+}, sulfide, and ammonia are stable. If oxygen can take up electrons and reduce electron availability (oxidizing conditions), the stable oxidation states are Fe (III), SO_4^{2-}, and NO_3^-. Table 49 gives the order of utilization of principal electron acceptors in soils and the other redox characteristics.

Different soil types are characterized by their specific features in the development of the redox processes. Podzolic and soddy-podzolic soils within normal wetness have Eh in the range of 550-750 mV, whereas chernozems have it within the range of 400-600 mV. Grey semidesert soils have the lowest Eh values of about 200 mV.

Seasonal variability of soil temperature, aeration and wetness, as well as the peculiarities of soil microbiological processes determine the dynamics of the redox potential that is the redox regime of the soil. This regime may be defined as the ratio between the oxidation and reduction processes in soil profile during the annual cycle of soil formation (*Soil Science Textbook* edited by I.S. Kaurichev). The redox regime of the soils may be of four types:

Table 49. Order of Utilization of Principal Electron Acceptors in Soils, Equilibrium Potentials of These Half-Reactions at pH 7, and Measured Potentials of these Reactions in Soils

Reaction	Eh at pH 7 (v)	Measured Redox Potentials in soils (v)
O_2 Disappearance $1/2O_2 + 2e + 2H^+ = H_2O$	0.82	0.6 to 0.4
NO_3^- Disappearance $NO_3^- + 2e + 2H^+ = NO_2^- + H_2O$	0.54	0.5 to 0.2
Mn^{2+} Formation $MnO_2 + 2y + 4H^+ = Mn^{2+} + 2H_2O$	0.40	0.4 to 0.2
Fe^{2+} Formation $FeOOH + y + 3H^+ = Fe^{2+} + 2H_2O$	0.17	0.3 to 0.1
CH_4 Formation (example of fermentation) $(CH_2O)_n = n/2CO_2 + n/2CH_4$	–	-0.15 to -0.22

1) soils with the absolute dominance of oxidation processes, such soils may be represented by the soils with low water table, such as chernozems, chestnut soils, grey-brown, brown semidesert soils, greyzems of arid subtropics, grey cinnamonic soils, etc.;

2) soils with the dominant oxidation conditions with possible manifestation of reduction ones in certain humid years or wet seasons. These are the automorphic (not affected by the ground water table or capillary fringe) podzolic, soddy-podzolic, grey forest and brown forest soils, oxisols of humid subtropics, etc.;

3) soils with contrastically changeable redox regime (semihydromorphic soils in various geographic zones). Such soils may have temporary excessive wetness and develop under the influence of capillary fringe. Temporary excessive wetness may alternate with soil dryness, so that the reduced forms of iron and manganese may be oxidized anew endowing soil horizons with the characteristic ochre spots of seasonal gleying. Such soil geneses are encountered among the podzolic, soddy-podzolic, and podzolized soils, as well as among the solots and solonetzic soil types;

4) soils with stable dominance of reduction processes represented by peat and muck soils as well as hydromorphic (wet) solonchaks.

Redox conditions in soils may vary widely over short distances. In aerobic soils the interior of soil aggregates may be anaerobic. In wet soils, only the largest pores are open to gas diffusion from the atmosphere. Although the oxygen supply may be adequate for aerobic plants and microbial activity along the walls of large pores, the bulk of the soil is likely to be oxygen deficient.

Redox conditions in saturated and flooded soils are more homogenous. But flooded soils with a good supply of readily decomposable OM may become depleted of oxygen within a day. Anaerobic and facultative microbes multiply and continue the decomposition process. In the absence of oxygen, other electron acceptors begin to function. Nitrate is reduced first, followed by manganic compounds, ferric compounds, sulfate and, finally, sulfide. The formation of S^{2-} (sulfide) occurs in a strongly reduced environment resulting in the formation of FeS_2 and H_2S. The subsequent artificial drainage of such soils may turn them into **acid sulfate soils,** not suitable for growing anything in them. While trying to regulate the redox potential in automorphic and semihydromophic soils with more or less tolerable aeration important worth while to remember that the optimum conditions for the nitrification are in the range of Eh within 250-500 mV.

Questions & Assignments

1. What is soil solution and how can it be separated from the soil?
2. What do you know on the composition, concentration, and osmotic pressure of soil solution?
3. Give examples of oxidation and reduction processes occurring in soils.
4. Prove that soil is a real oxidation-reduction system.
5. What is the redox potential of the various soil types?
6. What is the redox of the soil?
7. What types of the soil redox regime are identified in Ukrainian Soil science?
8. How can the redox regimes be regulated?
9. What is rH_2 and how can one evaluate the redox regime of the soil using this index?
10. What redox conditions and pH value favor nitrification?

Chapter 23

Soil Productivity and its Evaluation

There exist some discrepancies between Ukrainian and international terms denoting soil fertility and productivity. In Ukraine (as in many other former Soviet republics the definition of soil fertility was associated with the name of V.R. Williams who wrote: “Soil fertility is its ability to meet the needs of plants in the elements of nutrition, water, supply their root systems with sufficient amount of air and heat and create a favorable physic-chemical environment for their normal activity”. In H.D. Froth’s *Fundamental of Soil Science* it is stated that “soil fertility is the status of a soil with respect to its ability to supply elements essential for plant growth without a toxic concentration of any element”. Soil productivity in the same textbook is defined as “the soil’s capacity to produce a certain yield of crops or other plants with optimum management”. Some early western textbooks on soils definied factors controlling the growth of higher plants as (1) nutrients, (2) water, (3) heat, (4) air, (5) light, and (6) mechanical support. It was stated that the soil was an agent in supplying these factors, either or in part. Plants growth depends upon a favorable combination of these factors. Soil productivity can be not greater than that allowed by the most limiting of these factors.

So, it is not difficult to come to conclusion that the soil fertility is a part of its productivity and it is necessary to speak about the potential productivity of soils measured by the characteristics of their properties and regimes. This potential productivity can be evaluated in some dimensionless units without the regard of definite crops or other plants and with such a regard. “These two approaches should not counteract” with each other. There may and should be both general and concrete evaluations of soil and land resources serving different purposes.

23.1. Classification of Soil Productivity

Though soil productivity and even its somewhat abstract manifestation – potential productivity – is a single entity equal by its value to a definite number of grade points, it is a tradition in Ukrainian soil science to differentiate it into several not less definite entities.

Thus, soil productivity may be natural and artificial (or man-made). **Natural productivity** is characteristic of virgin soils which never experienced any anthropogenic effects. **Artificial productivity** is that added to the natural one as a result of human activities like irrigation, drainage, fertilization, the use of chemical amendments, etc. In some cases it might be closer to the truth to say that there is no addition to but subtraction from the natural soil productivity if the cases are those of soil degradation. Soil productivity is also divided into **potential** and **effective**.

Potential productivity is measured by the characteristics of soil properties and regimes. The values of these characteristics will show what can be expected from the soil in terms of the essential crop yields. **Effective (or economic)** productivity is measured by the factual stable yields of essential (worthy to produce in a given locality) crop yields. To our mind, effective productivity does not coincide with an economic one. The same yields of a given crop may be obtained on different soils with different inputs of money and care. Economic productivity should be evaluated by purely economic indices of profitability. Inputs and outputs should be properly analyzed in relation to each other. The economic productivity of a chermozem (Mollisol) is much higher in relation to winter wheat than the economic productivity of a soddy-podzolic soil (Spodosol), though the yields of grain (outputs) may be the same. Inputs will be different.

23.2. Factors and Conditions of Soil Productivity

It is always stated in Ukrainian textbooks that soil productivity is something divine making soils different from their parent materials. But it is good to keep in mind that parent materials may also be endowed with considerable fertility and be productive. Parent material is one of the factors of soil productivity determining to what extent the resulting soil may be “pedigreed” or “mongrel”, or rather “highborn” or a “lowborn”. In Ukrainian soil science the factors of soil productivity are the following: (1) elements of plant nutrition; (2) water; (3) air, and partly (4) heat (or rather, “warmth”). The conditions of soil productivity include the combination of soil properties and regimes the interaction of which may determine the limits of the soil to supply the plants with the factors of their productivity. Some authors identify the following soil regimes: (1) thermal, (2) water-related, (3) plant nutrition related, (4) related to soil air, and aeration, (5) related to the redox soil condition, and (6) related to the toxic substances in soil. Some other authors add to this list such soil regimes as ‘physico-chemical”, “microbiological”, “biochemical”, “related to light in the surficial layer”, etc. There may be no end to it, and it is no wander, as the soil is an extra-complex system.

23.3. Models of Soil Productivity

Modeling is an important aspect of system approach to soil evaluation from whatever point of view under consideration (for example, for the soil fertility evaluation, or for the assessment of a given soil (mapping unit) for cherry orchards). In the aspect of soil system a model must help analyze its present condition and predict the possible outcomes of its specialized use.

From a philosophical point of view, a model is a reflection of a system. Such a reflection can be more or less adequate to reality. But it must allow to obtain additional information on the changes occurring in the system under the influence of both internal and external agents.

The soil productivity as a system is composed of two parts (1) **matrix (structural)**, and (2) **functional**. **The matrix part of soil productivity** is composed of more or less stable soil properties in relation to plants growth and development, like soil texture, soil depth, OM content, CEC, etc. **The functional part of soil productivity** is created by the set of soil regimes: water-related, aeration-related, salinity-related, heat (temperature) regime, soil reaction, etc. The characteristics of soil regimes are less stable than those of matrix soil properties. They can be controlled and changed in accordance with human needs and interests. So the simplest form of soil productivity model may be represented by a list of optimum soil characteristics pertaining to the matrix (stable soil properties) and to the functional (soil regime characteristics) components of the soil productivity system.

One man's meat is another man's poison. What may be optimum for one crop or plant, may not be such to the other. Soil fertility optimums are very important for plant growth and development. Whatever has been done in the world in this respect has been properly summed up and analyzed by Charles A. Black, a distinguished Professor Emeritus in the Department of Agronomy at Iowa State University in his book *Soil Fertility Evaluation and Control.* Throughout the book, crop responses are emphasized as the ultimate basis for soil fertility evaluation and control. Response curves allow to identify the optimum values of the content of macro-and microelements of plant nutrition in soils for definite crops. There certainly are some inexorable limitations even to the use of response curves for the purpose, as the factors of plant growth and development interact with each other in a complicated way. These curves, though, may be used for other numerous purposes, like, for example, evaluating the supplies of nutrients in soils, the degree of sufficiency of nutrients for crops, the quantities of nutrients needed to meet specific objectives, the value of soil pH alterations with liming, the value of different fertilizers aplications, the residual effects of fertilization, the relative value of different sources of a common nutrient, and the economics of fertilization and soil amendment.

To identify some optimum parameters of soil productivity in general and fertility in particular, a parabolic model has been extensively used by researchers of the Ukrainian Research Institute of Soil Science and Agrochemistry (Kharkiv). This approach was employed to evaluate the outcomes of field experiments on soils and fertilizers carried out throughout Ukraine. The response curves were approximated to parabolas which could approximately be derived from the first order differential equation:

$$d_y / d_x = K(X_{opt} - X) \tag{70}$$

where y is the yield of a crop with total nutrient quantity X (total amount in the soil plus added with a fertilizer) or only added nutrient quantity X_1, K is a constant, and X_{opt} is the optimum value of total or added nutrient quantity (X_{opt} and X_1 opt. respectively). Integration of the initial equation yields the working equation of a parabola:

$$y_x = -0{,}5KX^2 + KX_{opt} \times X + C \tag{71}$$

where C is the integration constant and y_x is a concrete value of a crop yield corresponding to the X amount of a nutrient (either total, or added). The value of y approaches maximum as the value of X approaches optimum.

23.4. Agroecological Method of Soil Evaluation for Crop Production

23.4.1. Urgency of the Problem

Soil fertility (and productivity) problem, according to C. Black (1993), still remains one of the most important humanitarian problems of the next ten decades – feeding all the people due to arrive. The problem of producing enough food lies principally in the developing countries, or the countries like Ukraine, which are drastically changing the way of their agricultural production. For both, developed and developing countries, the soil fertility problem is one of efficient management and improved precision in evaluating all factors that influence plant responses and in applying a correct treatment in order to obtain the desired results.

To consider the general stock of country's soil resources and to evaluate the soil units and land areas for the production of a given group of crops are two different tasks. Some procedures of the general soil and land evaluation, like that developed by the late Professor A.I. Siry in Ukraine, allow to take a panoramic view of soil and land resources in a given region as well as on a given farm. But these methodologies, even though quite worthy of existence and fruitful in resource evaluation, remain practically impotent when a need arises to identify the most suitable soils and plots for the production of a given crop or to classify them for their suitability for other specialized uses. What is good for the potatoes may not be so good for the sunflower crop. A panoramic view of trends may lack the details of constructiveness.

Research imperatives of crop productivity today should include soil impact in such a way as to be able to prevent the processes of social and economic transformation carried out by the pecuniary western pattern from being destructive to our common cradle – the environment. Importing new ideas and politics from abroad, we should also ask the crop what it needs and then look how it affects the soil and the bordering environments. There should be two approaches to soil and land evaluation by a specialist in soil science, agronomy, and economics: (1) general and (2) specialized. In any of these approaches it would be good not to tear off soil conditions from the climatic and other environmental ones. **Soil-climate conditions make an integrated system and it is, to our mind, better not to separate them from each other**. Such a system may serve the purposes of any further (say, economic) land assessment, and actually serve a foundation on which to build this assessment.

23.4.2. Evaluation of Soil Condition for Plant Growth

The Ukrainian Research Institute of Soil Science and Agrochemistry (URISSA) has been constructive in systemizing soil and climatic conditions of plant growth and development for nine main field crops: winter wheat, rye, barley, oat, maize, sugar beet, potato, sunflower, and flax. The researchers identified optimum, satisfactory (admissible) and unsatisfactory (nonadmissible) conditions for the growth of each kind of crop. The selection of parameters included the depth (cm) of a surface layer richest in the OM content (humusified layer), soil textural class, bulk density (g/cm^3), pH$_{H2O}$, OM (humus) content (%), the sum of effective (>10°C) air temperatures at germination, air temperature at the formation of spike or other stages of growth and development important for a crop under consideration, hydrothermal coefficient, depth (cm), of ground water table, ground water salinity (for the water table depth below 5 m. g/liter), the stocks of productive (available) moisture in the 0-20 cm layer of soil at the emergence of crop seedlings (mm), the same parameter in the 0-100 cm layer at the stage of flowering (mm), etc. Obviously that the indices of soil pollution with radioactive wastes, pesticides, and heavy metals, as well as soil infestation with weed seeds should also be added to this list.

The ecological demands ("normatives") were published in the form of a table. For a number of soil and nonsoil environmental parameters (pH, bulk density, air temperature, etc.), the upper and lower limits of satisfactoriness (tolerability) were stated. The evaluation of soils should be performed with the consideration of not only their own natural properties, but also with a view of factors necessary for the crops themselves. In the project entitled "Agricultural Land Use Planning Based on the Integral Evaluation of Land Suitability for Crop Production" the researchers (V.V. Medvedev, T.N. Laktionova, et al.) elaborated the methodology consisting of the following stages: (1) indices selection; (2) setting the rates of indices (rending among the three suitability levels: optimum, satisfactory, and unsatisfactory); (3) mapping spatial distribution of each index (parameter), and (4) integrating information and compiling of a map of land suitability for specific crops. The research allowed finally to compile nine integrated maps for nine essential crops.

23.4.3. Methodology of Soil (and Land) Value Indices Computation

Doing their own research, the authors of this book propose to apply the "ecologically biased" **desirability functions** of T.A. Grinchenko to the procedure of soil and land evaluation based on the needed for growing good agricultural crops.

Any parameter taken in its specific units can be transformed into grade points (0-100 points), designated here by y_1 100 points corresponding to the optimum value of a parameter. Unimodal desirability function looks like that:

$$y_1 = 100 \times \exp\left\{ -5 \left| \frac{X_i - X_{opt}}{X_{opt}} \right|^3 \right\} \qquad (72)$$

Here X_i is the factual parameter of a soil property in its specific unit and X_{opt} is the optimum value of a given parameter. Numbers – 5 and 3 just proved most suitable for the purpose, though some other coefficients may prove "better adjusted". But there must not be any "blind" use of the formula. If the optimum value of humus content in the conventional plow layer (0-20 cm) of soil for winter wheat is over 3.5%, and actual value is that of 4%, it means that this value should be estimated by 100 points and the need for computations disappears.

But if the factual humus content is 2%, y_{hum} is computed in the following way:

$$y_{hum} = 100 \times \exp\left\{-5\left|\frac{2.0-3.5}{3.5}\right|^3\right\} = 67 po\operatorname{int} s \tag{73}$$

Optimum value of soil pH for winter wheat lies within a range of 6.0-7.5. There are two unsatisfactory levels of a given parameter: pH < 5.6 and pH > 8.0. The factual value is 5.8. It is closer to 6.0 than to 7.5. Y_{pH} is, therefore, computed in accordance with the equation:

$$y_{hum} = 100 \times \exp\left\{-5\left|\frac{5.8-6.0}{6.0}\right|^3\right\} = 99.98 = 100 po\operatorname{int} s \tag{74}$$

The integrated grade point is computed from the set of individual *n* points as their geometric average value, as it was proposed by T.A. Grinchenko:

$$ISGI = \sqrt[n]{y_1 \times y_{2\times} \times \ldots \times y_n} \tag{75}$$

where ISGI stands for the INTEGRATED SOIL QUALITY INDEX for a given soil (mapping unit).

Any land area (plot) is made up of a set of soil units, like polypedons, or mapping units on a soil map (scale 1:10 000) of an individual farm in Ukraine. ISGI for a given area of land (a field or a crop rotation) can be computed as a weighted average value by the area of a soil unit:

$$ISGI\ for\ \ a\ \ plot\ \ of\ \ land\ = (ISGI)_1 \times S_1 + \ldots + (ISGI)_n \times S_n / S_1 + \ldots + S_n \tag{76}$$

here $(ISGI)_1 \ldots (ISGI)_n$ are integrated soil quality indices of individual soil units on the plot, and $S_1 \ldots S_n$ are the areas (ha) of each of the soil units on the plot.

The weighed averages obtained in this way could be modified by multiplying them by respective modifying coefficients, which, among other aspects of soil productivity, could reflect economic conditions of land use, such as field area configuration, and the distance to the nearest marketing areas or processing factories.

23.4.4. Interpretting Results

Owing to the research done by V.V. Medvedyev, T.N. Laktionova, et al., the maps of land suitability contain information in which part of Ukraine the conditions are optimum, satisfactory, or unsatisfactory for the production of each kind of crop. The Ukrainian territory is divided into regions, with regard to land quality for agricultural crop production. The application of this approach at a countrywide or a regime scale makes it possible to improve the processes of plot selection. Rational land use in agriculture results in more meaningful resources consuming and the stabilization of the environment so necessary for the sustainable development. Long-term and versatile soil investigation within Ukraine has helped and will continue to help collect data which could be easily processed in the up-to-date programs with electronic cartography techniques (Fig. 95).

Fig. 95. Soil Fertility in Ukraine

Geo-information system (GIS) data bases will be used for a number of purposes, among which providing the users with information on soil properties and lands conditions of will not be the least.

Still, the evaluation of soil properties and lands in Ukraine remains a difficult problem. Soil (and land) quality should become an important factor in determining the price of a plot. It is not yet thoroughly taken into account in yield programming and crop production systems.

Questions & Assignments

1. What is soil fertility? In what relation is it to soil productivity?
2. Compare the factors of soil fertility in Sokolowski's hexagon with the factors of plant growth (soil productivity) in western textbooks of Soil Science.
3. Traditional classification of soil productivity still in use in Ukraine. What are in your opinion its strong and weak points?
4. What factors and conditions of soil productivity do you know?
5. What are the models of soil productivity? Which component parts do they usually consist of?
6. Name soil regimes known to you? By what parameters are they characterized?
7. Speak on the approaches to soil productivity evaluation in Ukraine.
8. What is desirability function and how it can be used for soil quality evaluation?

GENESIS, CLASSIFICATION, GEOGRAPHY, AND AGRICULTURAL USE OF SOILS

The study of soil genesis and development includes the study of the soil-forming factors, both natural and anthropogenic, which were considered in the previous chapters of Part One. So this part of the textbook will stress the specific features of soil formation in connection with soil typology and geographic distribution. Much attention will be devoted to the genetic ecologic-substantive soil classification proposed in Ukraine by M.I. Polupan et al. (2005). Formerly Ukrainian soil science was using the classification proposed by the Dokuchayev Soil Research Institute (Moscow) in 1962. The newest Ukrainian classification is the freshest if not the best approach to soil diagnostics and nomenclature. To this day, Ukrainian classification still suffers from the absence of unified and agreed upon approaches. The Ukrainian soil scientists will, no doubt, make efforts for the improvement of a situation in this field, taking into account the other countries' achievements.

Chapter 24

Soil Classification, Nomenclature, and Diagnostics

24.1. General Notions

Classification schemes of natural objects are designated to organize knowledge, so that the properties and relationships of these objects may be remembered and understood for some special purpose. Most soil classification schemes have been developed on a national basis. To date, the most suitable for international purposes is the *Soil Classification and Nomenclature* developed by the FAO organization under the United Nations.

Soils vary from place to place in many profile characteristics. Each soil, as has been already stated in this textbook, is a three-dimensional individuum (pedon) which should be identified and classified in the field. Polypedons in Ukrainian soil science coincide with mapping units in large-scale (1:10 000) soil survey. Each pedon may vary in size, say, from 1 to 10 square meters. *Pedon* in Greek means *ground*. It is the smallest volume that can be called a soil. Because of its very small size, a pedon obviously cannot be used as a basic unit for a workable field soil classification. A group of pedons closely associated in the field and similar in their essential properties are of a sufficient size to serve as a basic classification unit. Such a group approximates what is called a soil series in old and new comprehensive American systems of soil classification. Ukrainian systems were and remain very much different. Any of the Ukrainian systems is genetic, geographic, and historic in its essence, based on the concept of Dokuchaev soil science. Classification systems based on the inherent properties of the objects classified are called *natural* or *taxonomic systems*.

The system of soil classification inherited by our country from the former Soviet Union included the following taxonomic units: 1) class, 2) type, 3) subtype, 4) genus, 5) species, 6) variety, and 7) lithologic series. The soil of Ukraine has been divided into two classes: 1) zonal soils (bioclimatogenic soils) and 2) azonal soils (biolothohydrogenic soils). Roughly, the characteristics of zonal soils are determined primarily by the climate and vegetation. As the name *zonal* indicates, these soils are of such a wide expanse as to be more or less regional in extent. Azonal soils, in spite of climate and vegetation, reflect the influence of some local conditions such as poor drainage or alkali salts, or lack proper horizon differentiation. Most soils developed in recent alluvial deposits are azonal ones. Much remains to be elucidated done in this field of Ukrainian soil science.

A soil type has been viewed as the basic, fundamental taxonomic unit. The soils belonging to the same type were formed under typically uniform plant associations, in parent materials of similar mineralogical composition and endowed with a similar type of soil profile, like level of soil productivity and some type-specific set of soil management practices.

Soil subtypes were identified to allow deciphering the content of a soil type. Subtypes included the soils that, in addition to type-specific, possessed some properties of the other soil types. Subtypes were viewed as the intermediate chains between the types which reflect the nondiscrete nature of soil cover.

Soil geneses were identified within the subtypes by the set of genetic (adhering to origin) soil properties brought about by the specific features of parent materials, ground water composition, or the properties preserved from the preceding stage of soil formation. The classification system took into account the presence or absence of carbonates, soil acidity, reduction processes development (gleying), depth of salinity or sodicity in soil profile, presence of claypans, skeletal fragments or the features acquired by the soil as a result of management.

Soil species characterized the extent of the development of typical, subtypical, genial, and lithologic soil properties. Species took into account the depth of humus-containing horizons, organic matter content, and the extent of soil development or degradation (erosion, salinity, sodicity, carbonate level, etc.).

Soil variety reflected soil-texture in the former classification.

Soil lithologic series were determined by the character of parent material and occasionally also by the material immediately underlying the horizons of soil profile.

24.2. Principles of Genetic Ecologico-Substantive Soil Classification

The principles discussed here were proposed by M.I. Polupan et al. (2005). They include the following principles: 1) zonality, 2) genesis, 3) hierarchy, 4) substantivity, 5) inheritability, and 6) reasonable purposefulness.

Zonality principle in soil classification. Soil as a natural body and the object of human activity, the basic means of agricultural production cannot be isolated from the geographic conditions of the formation. Functional dependence of soil genesis and properties on the factors of soil formation constitutes the basic law of soil science which reflects the unity between the soil and its environment. This law had been introduced to soil science as the concept of soil zones which was fully employed in the development of soil typology. Such a concept is not only theoretically but also practically important for the correct placement and efficient management in agriculture.

Soil genesis principle is traditionally reflected through the evaluation of soil profile. In the existent soil classifications such an evaluation carried out by the procedure of morphological analysis and soil profile description. The Ukrainian Research Institute of Soil Science and Agrochemistry holds priority in the identification of correspondence of soil types to the modes of soil organic matter content and its distribution in soil profiles. Soil organic matter content is practically constant for a given soil profile. But the absolute values of soil organic matter content depend on soil texture. To make soil formation typologically comparable, the experts of the Institute proposed to employ the coefficient of Humus Accumulation in Soil Profile (CHASP), which is the ratio between the contents of soil humus and physical clay fraction (< 0.01 mm) in soil profile. The values of the CHASP are nearly constant for definite soil types (Table 50).

Table 50. Soil Types in Dependence on the CHASP in Various Natural Zones of Ukraine (M.I. Polupan, et al., 2005)

Zone	CHASP	Soil Type
Forest	0.020-0.040	Soddy-Podzolic Soil
	0.040-0.070	Podzolized Soddy Soil
	0.080-0.110	Gleyed Soddy Soil
Forest-Steppe	0.023-0.031	Light Grey Forest Soil
	0.031-0.040	Grey Forest Soil
	0.040-0.050	Dark Grey Podzolized Soil
	0.051-0.070	Podzolized Chernozem
	0.045-0.100	Typical Chernozem
	0.090-0.125	Meadow Typical Chernozem
Northern Steppe	0.055-0.065	Ordinary Chernozem
	0.065-0.085	Meadow Ordinary Chernozem
Southern Steppe	0.045-0.055	Southern Chernozem
	0.055-0.065	Meadow Southern Chernozem
	0.034-0.045	Meadow Southern Chernozem Gleyed from the surface
Arid Steppe	0.035-0.045	Dark Chestnut Soil
	0.030-0.035	Sodicity-Afected Chestnut Soil
	0.020-0.030	Chestnut Sodic (solonetz) Soil

Hierarchy principle in soil classification is observed through a definite system of taxonomic units of different orders, each of which is characterized by its own quantitative parameter.

Practically, all soil classifications, beginning with those developed by V. Dokuchaev and M. Sibirtsev, declared their adherence to the **principle of substantivity** determined by the parameters of soil composition and properties. Soil is a system the properties of which are interconnected and subject to the consequences of cause-and-effect relations. By its very nature, soil formation is the transformation of organic matter and synthesis of humus. M.I. Polupan, et al. (2005) stress the importance of dependence identified by their research activities between hydrothermal condition in the May-September period, humus accumulation, and type of soil formation for a given soil textural class. They propose to view soil humus as an index of typologic and ecologic "soil memory". The impacts of such a memory are reflected in soil organic profile and humus content alongside with the system of soil genetic horizons. The specific features of soil humus accumulation through the corresponding parameters implement the principle of substantivity in each of the taxonomic orders.

The inheritability principle of soil classification does not allow any nonmotivated change in soil nomenclature which provides a betted understanding and effective use of typologic definitions by the users.

It is next to impossible to rapidly meet all the requirements of the above stated principles. Certain deviations are inadmissible, which justified the adoption of the principle of reasonable purpose fulness in the use of the criteria.

24.3. Taxonomic Units of Soil Classification

The latest publication of M.I. Polupan, et al. (2005) proposes the system of soil taxonomic units baced on the hierarchy principles. Soil classification makes use of the following taxonomic units:

$$type \to subtype \to genus \to species \to \mathrm{var}iant \to litho\,\mathrm{log}ic \quad series$$

24.3.1. Soil Type as the Basic Taxonomic Category

Soil type is a group of soils united by the common system of soil horizons and organic matter profile on the basis of the CHASP values and as a consequence of the likeness in the regimes and processes of soil formation caused by the relative uniformity **of bioclimatic conditions**. During the field soil survey the types are diagnosed by the constitution of soil profile and morphologic features of the genetic horizons. A more exact identification is made by the parameters of organic matter accumulation in soil profile and the value of the CHASP.

24.3.2. Soil Subtype as the Subzonal Taxonomic Unit

A subtype allows deciphering the essence of a type by the intensity of SOM accumulation in the upper part of soil profile. The intensity is determined by the character of bioclimatic conditions. It is diagnosed by the value of the Coefficient of Relative Humus Accumulation (CRHA). This coefficient is the ratio between the content of humus and that of physical clay in the 0-30 cm layer of soil related to 10% of the physical clay content. The ranges of the CRHA were worked out for each type of soil formation (Table 15.3). For any given subzone, the ranges of the CRHA for definite soil types do not coincide.

24.3.3. Soil Genus as the Textural Matrix of Subzonal Soil Formation

Soil texture is very important for its agronomic quality and classification. But in soil classifications now used in Ukraine soil texture unjustly occupies the lowest taxonomic level. Polupan et al. (2005) propose that soil texture determine soil genus. Absolute values of soil organic matter accumulation for any given intensity of this accumulation are determined by soil texture. To compute them, it is necessary to multiply the CRHA parameters of any type of soil formation by the content of physical clay and to divide the obtained value by 10. Soil differentiation according to genus is given in Table 51.

Table 51. Typologic Gradation of Soils on the Level of a Genus

Number	Soil Genus Determined by Texture	Physical Clay (< 0.01 mm) Content, %
1	Sand	0-5
2	Bound sand	5-10
3	Light loamy sand	11-15
4	Heavy loamy sand	16-20

Number	Soil Genus Determined by Texture	Physical Clay (< 0.01 mm) Content, %
5	Sandy light loam	21-25
6	Light loam	26-30
7	Light-Medium loam	31-35
8	Medium loam	36-40
9	Heavy Medium loam	41-45
10	Light Heavy loam	46-50
11	Heavy loam	51-55
12	Light Clay	56-60
13	Light Medium Clay	61-65
14	Medium Cay	66-70
15	Heavy Clay	71-75

24.3.4. Soil Species: the Energy of Soil Formation as Effected by Abiotic Agents

A soil species reflects the differentiation of soil formation by the extent of moisture supply, amount of cold-season precipitation and its utilization by the soil, the effects of landform factors, soluble salts, and the content of skeletal fractions of soil texture.

The special habits of soils depend on the climate features. Hydrothermal characteristics of the warm period of the year are characterized by the parameters of relative SOM accumulation on the levels of soil type and subtype, whereas a soil species reflects the absolute amount of SOM. Moisture accumulation during a cold season is very important for the energetics of soil.

Soil formation and agronomic potential of lands. It is determined by the absolute amount of precipitation and its consumption of moisture by the soil. M.I. Polupan et al. (2005) have found the connection between the length of a frosty period and the consumption of moisture by the soil. Six “facial” soil species were identified on this basis. The soil profile depth is the reflection of the amount of winter precipitation and the facial features of their consumption by the soil. This depth is within the range of 25-200 cm. Special differentiation of soils by the depth of soil profile is shown in Table 52.

Table 52. Soil Species Determined by the Depth of Soil Profile (M.I. Polupan, et al., 2005)

Number	Soil Profile Depth, cm	Soil Species
1	< 25	Shallow
2	25-45	Short
3	45-65	Average Not Deep of Depth
4	65-85	Average Not Deep of Depth
5	85-105	Moderately Deep
6	105-125	Deep
7	125-145	Excessively Deep
8	> 145	Extra Deep

Soil species are also identified by the extent of reduction processes (gleying) development. Thus non-gleyed soddy-podzolic soils have the CRHA within 0.50-0.65, slightly gleyed: 0.66-0.82, and gleyed: 0.83-0.95. For podzolized soddy soils these values respectively are: 0.96-1.25; 1.26-1.60, and 1.61-2.00. Light grey forest soils are differentiated into gleyed from the surface (CRHA = 0.57-0.60) and slightly gleyed from the surface (CRHA = 0.60-0.65). Slightly gleyed from the surface species are also identified among the grey forest soils (CRHA = 0.610.73), dark grey podzolized soils (CRHA = 0.70-0.79), and podzolized chernozems (CRHA = 0.82-0.94).

Table 53. Differentiation of some Soil Types into Subtypes by the Intensity of SOM Accumulation in Accordance with the CRHA Values (M.I.Polupan, et al, 2005)

Grades of SOM Accumulation	CRHA	Soils									
		Brown Forest	Soddy-Podzolic	Light Grey	Grey	Dark Grey	Typical Chernozem	Ordinary Chernozem	Southern Chernozem	Chestnut	Sodic chestnut
Extra	4.00-5.00	3.80-5.00									
Sub-extra	3.00-4.00										
Very High	2.00-3.00										
High	1.60-2.00										
Moderately High	1.26-1.60						1.25-1.35				
Very Good	1.12-1.20						1.12-1.20				
Good	0.98-1.00						0.98-1.10				
Moderately Good	0.90-0.97					0.86-0.95		0.90-0.97			
Average	0.80-0.89		0.83-0.96	0.63-0.70	0.74-0.81	0.70-0.85		0.80-0.89			
Moderately Weak	0.68-0.78		0.66-0.82	0.56-0.65	0.61-0.73	0.68-0.76		0.68-0.78			
Weak	0.55-0.66		0.50-0.65	0.48-0.55	0.58-0.62	0.63-0.72			0.55-0.66		
Low	0.45-0.54			0.42-0.51	0.49-0.58						
Very Low	0.35-0.43									0.35-0.45	
Ultra Low	0.25-0.34										0.28-0.34

Soil species differentiation on the slopes should also be taken into account. The soils on the slopes occupy up to 30-50% of the total area of different soil zones. Their formation is different from that of the plain land soils in the regimes of temperature and water supply. For most of the slope soils the conditions are more arid as moisture is partially lost by surface runoff and soil temperature on the so-called “warm” slopes of eastern, south-eastern, southern and south-western aspects which receive more solar radiation, is higher. Worse water supply causes the lessening of the depth of soil moistening and creats less favorable conditions for plant growth which results in the reduction of SOM content. The more arid the conditions on the slope, the shallower the profiles of the soil and the lower SOM content. For a long time, the soils on the slopes were just considered eroded. But such a name does not always reflect their genesis and the cause for the reduction of their fertility. The landscapes on the slopes form and discharge their surface runoff through the preferential lines of flow, so M.I. Polupan et al. (2005) saw the main cause of profile depth reduction on the slopes not in soil erosion but in the worsening of soil water supply. The channels formed along the lines of preferential water flow were called by them “the papillars of runoff” or “surface runoff papillars” (SRP). By the extent of aridization the soils on the slopes are divided into slightly-, moderately-, and severely xeromorphic species. Slightly xeromorphic soils are characterized by the reduction of SOM in the plow layer by 8-22% and the reduction of profile depth by 10-25% as compared with the corresponding plainland soil species. For the moderately xeromorphic and severely xeromorphic species, these reductions respectively are 22-35 and 25-50%, and 35-50 and 50-70%. Soil species are also identified by the extent of soil salinity, sodicity, and skeletal fraction content. Saline soils are divided into species acording to the depth of the upper saline horizon in the profile, salt contents, and soluble salts chemical composition as determined by the soil extract analysis.

Considering the depth of the upper saline horizon (cm), the soils are divided into:

solonchaky (with saline surface):	0-30
salinity-affected soils with high salinity table:	30-50
salinity-affected soils with medium salinity table:	50-100
salinity-affected soils with deep salinity table:	100-150
and those soils with deep salinity:	Over 150-200

The extent of sodicity is determined by the morphological features of the profile as well as by the percentage of exchangeable Na^+ of the CEC: slightly sodicity-affected: (2.6-3.5%), and moderately sodicity-affected: (3.6-6.0%). Table 54 demonstrates soil classification according to the extent and chemical composition of soil salinity.

Table 54. Soil Classification according to the Extent and Chemical Composition of Soil Salinity

Extent of Soil Salinity	Total Toxic Salt Content, %					
	Chemical Composition of Soluble Salts					
	Chloridic	Sulphate-Chloridic	Chloride-sulfatic	Sulfate or chloride-bicarbonatic	Soda-chloridic or chloride-sodic	Soda-sulfatic or sulfate-sodic
Nonsaline	< 0.03	< 0.05	< 0.1	< 0.15	< 0.1	< 0.15
Slightly saline	0.03-0.1	0.05-0.12	0.1-0.25	0.15-0.3	0.1-0.15	0.15-0.25
Moderately saline	0.1-0.3	0.12-0.35	0.25-0.5	0.3-0.6	0.15-0.3	0.25-0.35
Severely saline	0.3-0.6	0.35-0.7	0.5-0.9	0.6-1.4	0.3-0.5	0.35-0.6
Very severely saline and solonchaks	> 0.6	> 0.7	> 0.9	> 1.4	> 0.5	> 0.6

Skeletal soils were formed in the eluvium of sedimentary, metamorphic, and solid crystalline rocks. Their surface layer (0-30 cm) may contain 5 to 95% of skeletal particles. Downward their content increases. Diagnostic parameters of skeletal soil species are given in Table 55.

Table 55. Diagnostic Parameters of Skeletal Soil Species

Soil Name	% of Skeleton Content	Relative Content, % of		
		Stones	Rubble	Chert
Slightly Cherty	10-20	–	20-30	70-80
Cherty	20-30	–	30-40	60-70
Rubble-Cherty	30-50	< 10	30-40	50-60
Stone-Cherty	50-80	10-30	30-40	30-50
Rubble-Stony	> 80	40-50	30-40	20-30

Soil skeleton reduces the useful volume of the soil mass. All calculation pertaining to skeletal soils must be performed on the skeleton-free weight.

24.3.5. Soil Variant: Taxonomic Unit Reflecting Soil Modification by Anthropogenic Agents

The plowing of virgin soils changes their ecological conditions, intensity, and the trends of some soil-forming processes, causes the deterioration of soil structure but improves water and nutrient supply, intensifies microbiological (related to soil colloids and adsorbing complex) properties. Reclaimed soils acquire certain specific morphologic features compared to their virgin counterparts. “Soil variant” as a taxonomic unit reflects the modification of soil properties in the course of soil’s agricultural utilization, mainly as a plowland for crop production or other farming system. Soils it is proposed by M.I. Polupan et al. (2005) to identify the following: modal, cultured, eroded, irrigated, sodicity-affected (in the process of irrigation), drained, and excessively wet (as a result of irrigation). There may be some other variants too, as it is difficult to foresee all possible anthropogenic modifications.

Modal variant means that the soils did not yet undergo any visible changes. Cultured variant is diagnosed by the content of available phosphorus and potassium which increases as compared to initial soil condition characterized by unchanged or increased SOM content in the plow layer.

Eroded soils, according to the cited authors, are found mainly on the bottoms of the surface runoff papillars (SRP).

Irrigated soils with low water table did not substantially change their initial properties. They are more eluviated and, therefore, contain less salt, but the content of exchangeable sodium in them is increased remaining though below the threshold of sodicity. The physical properties of irrigated soils mainly deteriorate to different extent. These features are sufficient to diagnose soil species as irrigated.

Repeatedly sodicity-affected and sodic soils is the term used in Ukraine to denote the soils which acquired the features of sodicity as a result of irrigation. Exchangeable sodium content usually exceeds the threshold of soil sodicity. Slightly sodicity-affected soils contain 2.6-3.5% of exchangeable sodium of the CEC, whereas moderately sodicity-affected soil variants contain it within 3.6-6%.

Repeatedly (secondarily, as translated directly from the Ukrainian) wet soils also were formed as a result of irrigation and ground water table elevation caused by it. The authors suggest identifying slightly meadowed and meadowed soils, the former – with ground water table within 3-5 and the latter with that less than 3 meters from the surface of the soil.

Drained variants are identified on the drained areas.

Drained irrigated variants are separately identified, as their formation takes place under specific conditions. They have better conditions for the SOM accumulation and have a greater buffering capacity against the development of sodicity. Their overall properties are better in comparison with just drained variants.

The variants of reclaimed soils are very numerous as are the practices of reclamation. Not all of them can be mentioned here, but some reflect various forms of **soil pollution.**

As the pollutants and their sources are very numerous, so are the variants of polluted soils. Two very important soil variants in this respect are those polluted with industrial or farm wastes and those **polluted by nuclear wastes**.

Soils polluted by the radionuclides after the Chernobyl disaster are identified by the density of soil pollution with ^{137}Cs and ^{90}Sr, expressed in Ki per square kilometer. **Slightly polluted soils** have these values respectively below 0.1 and below 0.02 Ki/km^2, whereas **very severely polluted** variants may be characterized by the values of over 15 and 3 Ki/km^2 respectively for ^{137}Cs and ^{90}Sr.

24.3.6. Lithological Series: Soil Identification based on the Parent Material Genesis

As the name implies, to identify a lithologic series it is necessary to dispose of certain information pertaining to parent material. By lithologic series, the soils may be loess, loess-like, ancient alluvial, clayey, sandy, limestone, granitic, etc. To our mind, the names of lothologic series are a failure of the authors of a given approach. Would not it be simpler and more to the point to state that the soil was formed **in** or **from** the loess, alluvial sand, glacio-fluvial loam, etc?

Ukrainian soil classification proposed by M.I. Polupan, et al. (2005) is given in Appendix Three of the book.

24.4. Diagnostics of Soil Ecologico-Substantive Status

The constitution of soil profile is determined by the sequence from the surface downwards or the genetic soil horizons. To diagnose the soil in the field one has primarily to describe soil horizons by their **morphologic features** which are ecologic and functional external features of the soil. Each genetic horizon since the Dokuchaev times is designated with a symbol.

24.4.1. Soil Horizons and Their Symbols

Soil horizons are genetically interconnected horizontal layers in the soil profile which differ from the initial parent material by morphologic features, as they were transformed by the processes of soil formation and endowed with functional properties.

Peaty (T) horizons contain over 70% of plant residues (those of trees, grasses, mosses, and lichens) at various stages of decomposition.

Peaty-humic (TH) horizons are composed of well decomposed and partly humified plant residues which are difficult to identify morphologically. They are dark-colored, of dusty, grainy, or crumby structure.

Peaty-mineral (TC) horizons are composed of the finely grained mineral and charred organic particles (even the smallest particles can be seen with an unaided eye). They are ash-like and hydrophobic. The wind can easily blow them up and carry away. Such horizons are found on overdrained peats.

Organic accumulative (H_0) horizons lie on the surface of the soil. They may be represented by forest litter or “steppe wool”.

Soddy or just sod (H_d) horizons are by a half or more of their mass composed of the living and dead roots of grasses.

Humus (H) horizons, or the horizons of humus or humified organic matter accumulation, are uniformly saturated with the SOM which, intern, is closely united with the mineral part of the soil. Horizons colors may be grey, dark grey, dark brown, or brown. They are loose, friable and well-structured with grains and clods.

Eluvial (E) horizons are poor in OM and clay particles which were partially washed downwards. Horizons color may be whitish, light grey or pale yellow. By their structure they can be platy or scaly.

Illuvial (I) horizons are the opposites of eluvial ones. They are rich in clay particles and labile sesquioxides and, occasionally, in OM. Their colors are brownish red, dark brown, brown or dark grey. Illuvial horizons are compacted, hard, prismatic, nutty, or prismatic-nutty in structure.

Pseudofibrous (Pf) horizons are made up by the alteration of thin brown or reddish-brown compacted layers (pseudofibres), 1 to 3 cm thick, with the layers of lightly colored sand, 5 to 10 centimeters in thickness.

Ortzandic (R) horizons are made up of hard layers of sand cemented by iron oxides mostly of hydrogenic or microbial origin. Usually they are red-colored, massive, and compacted.

Ortsteinic (R_g) horizons are rich in clay particles, sesquioxides, gels of silicic acid, hard, firm (ore-like), and reddish brown.

Solonetzic (sodic) (SE) horizons are intensively peptized (deflocculated), rich in labile clay, silica, and OM; their color is grey or black; their structure – prismatic or columnar with a characteristic “colloidal glittering” on the bases. They become structureless when wet, as well as sticky (adhesive) and swelling.

Gley (gleic, gleyed, etc.) (Gl) horizons may be mineral or organic-mineral, of dull grey, bluish, greenish, olive or of mottly coloring, structureless, formed in a reducing medium.

Marlic (M) horizons are composed from new formations of lime of hydrogenic origin (meadow marl). They contain in 28-59% of calcium and magnesium carbonates, are white or greish white, often with brown spots.

Transition (transitive) horizons retain some features of the two bordering horizons. They are designated by the symbols of these horizons (HE, EI, Hp, Ph, etc.).

Parent material (soil forming rock) (P), as the name implies, is geologic rock from which the soil was formed.

Underlying rock (material) (D) lies below parent materials.

Buried soil (Fs) is a fossil soil or the soil excluded from recent soil formation being covered up by some old or recent deposits.

24.4.2. Morphological Features of Soil Horizons

Soil color reflects lithologic and chemical composition of the horizon, as well as the quality of its organic matter. Wet soils are darker than dry ones. The state of soil wetness must be indicated before the indication of soil color. The illumination of the profile must be uniform all over, as shading makes the soil look darker. If the color is not uniform it is necessary to state the character of the dominant and modifying color. Special attention should be paid to spottiness. By the presence of spots, E.A. Kornblum, etc. et al. (1982) proposed the following gradation:

0 – soil color is uniform (no spots identified);
1 – slightly spotty (the spots are rare and not easily seen);
2 – spotty (the spots are well noticeable within 5-15 cm from each other); and
3 – very spotty (the spots are numerous, no more than 5 cm from each other).

The character of spots should also be stated, as they may be more or less distinct or diffusive in outline. Ukrainian soil diagnostics should start as soon as possible determining soil colors by matching the color of a soil sample with color chips in a Munsell soil-color book. The book consists of pages, each having color ships arranged systematically according to their **hue, value,** and **chroma**, the three variables that combine to give colors. **Hue** refers to the dominant wave length, or color of the light. **Value,** sometimes color brilliance (відтінок), refers to the quantity of light. It increases from dark to light colors. **Chroma** is the relative purity of the dominant wave length of the light. The three properties are always given in the order following: hue, value, and chroma. In the notation, 10 YR 6/4, 10 YR is the hue, 6 is the value, and 4 is the chroma. The latter is a light-yellowish brown.

The specter of colors to characterize soil horizons in Ukraine is so far very wide and subjective, therefore, it can be recommended to the reader to refer special instruction literature.

Soil wetness gradation accepted in Ukraine is as follows:

0 – dry (the presence of moisture is not felt by land; the soil does not become lighter on drying but it darkens with the addition of water; soil surface yields some dust to the wind);

1 – fresh (the soil sample is moist enough to slightly cool the hand, but the soil does not smear the hand and becomes of lighter color on drying and darker with addition of water);

2 – moist (the moisture is distinctly felt by the hand and the sample does not become darker with addition of water);

3 – very moist (the soil turns into a thick sort of putty on kneading, but water does not exude from it);

4 – wet when (squeezed in hand, the soil exudes water).

Soil texture is determined in the field by the above-described field procedure. Field method is insufficient to characterize soil texture. All kinds of soil survey need a particle-size analysis conducted in the laboratory.

Soil structure is determined in accordance with traditional S.O. Zakharov's classification. But the extent of soil aggregation should be an additional index to characterize soil structure:

0 – structureless soil (no signs of soil aggregation; natural cleavage lines are indistinct);

1 – weakly structured soil (the peds are not firm but they are visible in situ);

2 – moderately structured soil (the peds are distinctly and firmly formed but not easily identifiable in undisturbed soil mass);

3 – well structured soil (the peds are hard, distinctly identifiable in undisturbed soil, weakly united with each other and remain "whole" in a slightly disturbed soil).

Soil pores and cracks are characterized by their size. They maybe within and between the aggregates. Soil horizons may be nonporous if the number of pores does not exceed 5 per 100 cm^2, slightly porous (6-50), porous (51-200) and very porous (> 200). The pores (cracks) between the aggregates are classified by their thickness into thin (1-3 mm), average (3-10 mm), and large (> 10 mm) ones.

Soil hardness (penetrometer resistance) can be estimated in the field by the effortlessness with which the knife enters the soil in horizontal plane:

1. Loose soil – the knife easily penetrates it to the very handle;
2. Slightly compacted soil – it requires some effort to push the knife to the handle;
3. Compact – it is impossible to prod the soil with the knife to the handle even with considerable effort;

4. Very compact – the knife can be pushed into only a few millimeters of the soil. If there is a possibility to determine a bulk density of the soil, its hardness can be estimated according to Table 56.

Table 56. Soil Hardness as Measured by It's Bulk Density (g/cm^3)

Hardness	Plow layer	Subplow layer
Loose	< 1.1	< 1.2
Slightly compacted	1.1-1.3	1.2-1.4
Compact	1.3-1.4	1.4-1.5
Very compact	> 1.4	> 1.5

Sands and loamy sands usually have the bulk density within 1.50-1.65 g/cm^3.

Biological elements in soil are represented by the live roots and dead residues as well as by their parts and other morphological elements of organic origin. The number of roots in the soil is estimated according to the following scale:

0 – no roots;
1 – rare roots (1-5 items per 100 cm^2);
2 – frequent roots (5-50 items per 100 cm^2);
3 – numerous roots (over 50 items per 100 cm^2).

The roots are also characterized by their thickness in mm:

1 – very thin (< 1);
2 – thin (1-3);
3 – average (3-5);
4 – thick (> 5).

While describing organic horizons (forest litter, steppe wool, sod and peat) it is necessary to determine the extent of organic material decomposition according to the following scale:

1) slightly decomposed: plant residues preserved the form of a living plant but changed the color and character of surface;

2) moderately decomposed: plant residues changed their shape, color, and the character of surface. Some parts are still easily discernible;

3) well decomposed: plant residues lost their initial shape; only some fragments of changed plant tissues can be identified;

4) humified: plant residues are completely decomposed and transformed into dark colored mass which smears the hand. It is impossible to identify plant tissues with the unaided eye;

5) mineralized: plant residues are finely ground and charred and have a characteristic “ashy” look.

Botanical composition of plant residues, if possible, should also be determined or, at least, the identification of their origin (whether the residues are those of trees, grasses, or mosses).

Zoogenic elements in soils are represented mainly by animal remains: bones, shells, tusks, teeth, claws, scales, horns, feathers, eggshells, etc. Living animals, except earthworms, are rarely found. The quantity of zoogenic elements should always be stated in profile descriptions. Morphological elements related to plant and animal life cycles include root pores, rhizomes, excrements (earth warm casts), structural grains from ants, crotovinas, etc.

Segregated morphological elements are the result of soil formation. They differ from the bulk of the soil in color, consistency, chemical, and mineralogical composition. These may be the crystallized forms of soluble salts, carbonates, gypsum, ironous and iron-manganese concretions, amorphous silica dusting, films of clays or other colloids or cutnes. When describing these elements, it is necessary to indicate their form, color, hardness, and number per 100 cm^2.

Fragments of solid rocks and inclusions are characterized by their number, shape, size, the extent of weathering, etc.

The number of fragments may be:

1 – rare (1-5%);
2 – frequent (6-25%);
3 – numerous (26-50%);
4 – dominant (> 50% by volume).

Inclusions are solid morphologic elements that found their way to the soil as a result of human activity. These may be the bits of glass, porcelain, fragments of rocks rather different from the bulk of the soil mass, etc.

Their number may be small (1-5 items per 2-3 cubic decimeters) or large (over 5 items).

The majority of inclusions do not have any relation to the soil formation. They are mainly represented by the fragments of clastic rocks (Table 57).

Table 57. Soil Inclusions (according to I.S. Mikhailov, 1982)

Diameter, cm	Angular fragments	Smooth fragments
0.1-1	Chert	Gravel
1-10	Rubble	Pebbles
10-100 and more	Stones	Boulders

Rock inclusions may be unweathered (with fresh fracture surface), slightly weathered (outwardly firm but breaking to pieces with films on fracture surface), moderately weathered (friable, disintegrating into grains when pressed by fingers), and very weathered (can be cut with a knife, and it is difficult to extract them from the soil).

Borders between soil horizons depend upon the character of transition from the upper to lower horizons of soil profile. The transition may be sharp (clearly distinct), when the changes in horizon features occur within up to 3 cm in thickness, clear (when they occur within 3-5 cm), and gradual (when they occur in the thickness of over 5 cm. Border lines may be even, wavy, tongue-like, etc. The sequence of horizon alterations is of a regular character and caused by the type and intensity of soil formation, for example:

Typical chernozem: $H_{/k\ 40-50} + H_{pk60-80} + PH_{k80-140} + P_{hk140-200} + P_{k}$;

Light grey forest soil: $HE_{10-15} + E_{h30-40} + I_{h40-50} + I_{80-110} + I_{p110-130} + P_{k}$;

Chestnut sodic soil: $HE_{d0-7(15)} + E_{h8-15(20)} + I_{h16(21)-24(30)} + PI_{h/k25(31)-45(52)} + P_{k(h)46(53)-55(60)} + P_{k}S_{55(60)-170}$.

Every zonal soil has its own parameter of humus accumulation in the profile (SHASP), which is within 0.075-0.100 for typical chernozem, 0.023-0.031 for light grey forest soil, and 0.020-0.030 for chestnut sodic soil.

Questions & Assignments

1. Which soil characteristics belong to its morphological features?
2. Give examples of soil horizons and their symbols.
3. Explain the origin of whitish or light grey color of eluvial horizon.
4. Which diagnostic features are typical of illuvial horizons?
5. Give a list of the essential morphological features of soil horizons.
6. What agents does the soil color depend in the field?
7. What is soil type? Subtype? Genus? Species?
8. How many soil types are identified in Ukrainian soil classification proposed by M.I. Polupan, et al. (2005)?
9. How does a soil variant reflect the anthropogenic factor of soil formation?
10. What is the difference between new formations and inclusions?

Chapter 25

Soils of the Ukrainian Polissya (Forest Zone)

Polissya is a lowland with glaciofluvial sands, moraines, and alluvial deposits as parent materials of soils. It occupies the basin of the Prypyat, and the middle reaches of the Dnipro and the Desna. Ukrainian Polissya is a south-western part of the area of mixed forest; it belongs to the Bilorussian part of the southern subzone of the forest Eurasian zone. Southern subzone is the area of soddy-podzolic soils. In the north Ukrainian Polissya has a boundary with Bilorussian Polissya, whereas in the north-east it is neighboring with Bryansk Polissya. The southern boundary of the Polissya zone outlines an unbroken area of loess plate and of the Forest-Steppe zone.

Ukrainian Polissya is defined as a zone of pine and mixed forests with soddy-podzolic modal and gleyed soils. Its total area is 11.768 thousand hectares (19.5% of the Ukrainian territory). It includes nearly entire Volyn, as well as Rivne, Zhitomir, and Chernigiv oblasts, and northern districts of Lviv, Ternopil, Kyiv, and Sumy oblasts. Western, northern, and eastern borders of Ukrainian Polissya coincide with the state borderline of Ukraine, whereas its southern boundary in the right-bank part outlines the outcrops of the Volyno-Podilska and Prydniprovska uplands and in the left-bank part it runs along the line: Kyiv – Nizhyn – Baturin – Krolevets – Glukhiv.

Plowland takes up 49.6% of the zone's total area. There are numerous meadows, forests, shrubberies, and swamps in Ukrainian Polissya. Soil cover is rather complex with a great number of mapping units.

25.1. Factors of Soil Formation

Climate in the Polissya zone is moderately continental with warm humid summer and mild winter. The western part experiences the influence of humid Atlantic air, whereas the eastern part – that of continental Arctic air. The climate becomes more continental in the eastern direction. The total annual solar radiation is within 380-420 kJoules per square centimeter. Radiation balance fluctuates within a narrow range of 167-176 kJoules per square centimeter. Mean annual temperatures are within 5-7°C, lowering in the eastern direction. The annual total of active temperatures exceeding 10°C is within the range of 2,400-2,600°C, and the number of days with mean temperatures over 10°C is within 140-160 days. The rostless period lasts for 170-175 days, the snow cover period – up to 120 days.

Humidity is slightly excessive, as the annual precipitation reaches 550-650 mm, its main part (400-450 mm) occuring in the warm period of the year – by from April to October. Humidity coefficient fluctuates within 1.1-1.3. High ground water table is also one of the zone’s characteristic features. The excess of precipitation over the evaporation of water causes an infiltrative type of water regime, accompanied by swamping of soils in the depressions of topography. Such a phenomenon is especially widespread in the western provinces of the zone.

Topography. The territory of Ukrainian Polissya is a poorly drained plainland with well developed mesorelief land forms represented moraine hills and ridges, sand dunes, and mounds. In the Western Polissya, chalk deposits form large mounds and hills, but the lowlands between them are covered mainly by glaciofluvial and alluvial sands. Within the areas of the Ukrainian crystalline shield, the landscapes were formed mostly on glaciofluvial deposits and moraines. There are also some moraine hills, ridges, and kames (steep-sided, flat-topped hills of stratified drift).

Ridges and uplands have absolute altitudes in the range of 200-250 meters. The highest of them is the Slovechno-Ovruch ridge (320 meters above the sea level). Watershed topography is hilly and wavy, typical of the areas of glacial outwash and till alternating with each other. The lowest areas covered with glacial outwash favored the formation of peat and muck soils.

On the terraces, ridge sandy plains dominate. Loess islands jut up on sandy plains. They are severally eroded and cut with gullies. Subsurface runoff into Polissya lowland from the surrounding uplands feeds large and small rivers. Poor drainage and high watertable combine in the formation of swamps.

Parent materials in the Polissya zone are mainly represented by glacial, glaciofluvial, and ancient alluvial deposits. Loess and other materials can also be found. Moraine occupies the upper elements of topography, forming “islands”. It can also occur on the slopes, being mostly of loamy-sandy or light-loamy texture, red-brown or yellow-brownish in color, and containing boulders, pebbles, gravel, as well as angular inclusions of solid rocks. The most fertile soddy-podzolic soils form in moraine due to the richness of its mineralogical and chemical composition compared to glacial till or alluvial deposits. Some moraines can contain the fragments of limestonic rocks which impedes the development of podzolic soil formation. In some places, the moraines are heavy-textured and with low water infiltration rate, causing the reduction processes (gleying) in the upper portions of soil profiles. Fluvioglacial deposits (sands and loamy sands) dominate in the Polissya zone. Their sand particles are yellow colored (iron oxide coated) and contain some grains of gravel and even pebbles. They are slightly richer than grey-colored alluvial sands as to soil fertility.

Floodplains and terraces of the river valleys are covered with recent and ancient alluvial deposits. Soils formed in them are poor and infertile. Some of them are so light-textured that their water-holding capacity is too low for field crops. Pine forests, though, feel grow well comfortable on such soils and subsoils.

Generally, parent materials of the Polissian zone are light-textured and free of carbonates (with the exception of chalk marl in the west and loess on the "loess islands" elsewhere).

Vegetation is considered by many scientists as a leading factor of soil formation. The types of natural vegetation in the Polissya zone are represented by forest, swamp, and meadow associations. The forests were destroyed to a considerable extent. Pine forests occupy 57% of the total forest area, oak ones – 21, birch – 10, and alder – 6. Grass vegetation is only slightly developed in coniferous forests and does not effect soil formation to any appreciable extent.

The greater part of the zone is covered by mixed forests. Pines, which form the upper canopy, are mixed with oaks and other deciduous trees which form the lower canopy. Spruce forests may be encountered only in the northern part of the zone. Birch and alder forests occupy only small areas. The shrubs of hazelnut, hawthorn, and small crab apple trees, wild pear trees, and elder may form a lower canopy in the oak forests ("dibrovas" in Ukrainian). The grasses in such forests are better developed than in pine ones, and are represented by stellaria, buttercup, blackberry, strawberry, plantain, wood betony and other species of the Labiaceae family.

Meadow vegetation covers the lowered areas with a high watertable preventing the growth of trees. Such areas are called *forest glades*. But well-developed and quite productive meadow grasses grow within the floodplain meadows and grasslands.

Fifty percent of the area of Ukrainian swampland is located in the Polissya zone. Periodically flooded meadows are occupying vast areas in the floodplains of Polissian rivers. Lowland swamps dominate in the zone, upland ones occupying only 5% of the total area of swamps. Forests proper occupy only 29% of the zone's area. The main forest formations are pine, hornbeam, oak-pine, broad-leaved-hornbeam and alder in the western and central right-bank Polissya, pine and oak-pine- in the left-bank Polissya. Dryland meadows are found on the forest clearings or on the areas of reduced forests. The orchard grass, rough-stalk meadow grass, lovegrass, creeping fescue and clover are dominant species in the herb associations of such meadows. High water table meadows with areas flooded for more than 20 days are covered by more water-loving species like meadow foxtail, wheat grass, meadow grass, carex, buttercups, sorrel, etc. The vegetative cover of swamps is very biodiverse, including carex, reeds, cattails, mosses, etc.

Not too demanding to water, psammophytes[*] occupy the sand ridges and danes. These are mainly cereal grasses.

Agricultural crops grown in the Polissya zone include and entend on large areas rye, flax, potato, winter wheat, oat, vegetables, and lupin.

25.2. Soil Genesis in the Polissya Zone

The essential soil-forming processes in the Polissya zone include podzol formation, sod formation, and swamping (which is the formation of bog and peat of soils).

The podzolic process of soil formation takes place most intensively under the canopy of forest vegetation with infiltrative type of water regime (humidity coefficient exceeding 1) in parent materials free from calcium and magnesium carbonates (both initially free or leached), and is accompanied by the formation of organic acids by the fungal microflora decomposing forest litter together with anaerobic bacteria. Organic acids attack soil minerals and cause their destruction. The products of mineral decomposition in the form of molecular and colloidal solutions leach downward by the percolating activity of water and may reach parent materials and ground water. Clay particles may also move from upper to lower parts of the soil profile without decomposition or disintegration, the process being known as *lessivage.* The platy particles of silicate clays keep their shape and are specifically (vertically) oriented in the illuvial horizons, as can be detected by a microscope inspection. Crude forest litter containing cellulose, lignin, and tannins is hidden from the direct sun rays and, being under the influence of sufficient moisture, it favors the growth of fungi. Fungi, actinomycetes, and to a lesser extent, anaerobic bacteria are the major decomposers of the residues in the forest. They exude specific

---

[*] psammophytes are sand-loving herb species

enzymes into soil environment and decompose the components of forest litter with the evolution of CO_2 and formation of ammonium and phosphate salts as well as specific organic substances of polymeric acidic nature, mostly fulvic acids.

Soil solution dissolves and hydrolyzes soil minerals. Alaminosilicates are decomposed mainly by the living organisms. Thus, orthoclase and microcline may be decomposed by bacteria. Silica $(SoO_2 \cdot nH_2O)$ precipitates in the form of chalcedony or quartz, or remains in the soil solution. The soil becomes enriched with secondary silicate clays, hydrons silica, aluminum and iron hydroxides $(Al_2O_3 \cdot nH_2O, Fe_2O_3 \cdot nH_2O)$. Manganese hydroxides may also appear in the soil $(MnO_2 \cdot nH_2O)$. The infiltration of water through the soil transfers the soluble decomposition products from upper to lower soil horizons. The following horizons form under the layer of the forest litter (H_0): HE – humuso-eluvial, E – eluvial, I – illuvial, PI (P_i) – transitive to parent materials, and P – usually carbonate-free, or leached from carbonates, a parent material.

Fulvic acids accumulate mainly in the HE horizon where they are retained by the cations of Ca^{2+} and Mg^{2+}, which came from the decomposing forest litter and were retained by the adsorbing capacity of minerals. Oxides of iron and aluminum are retained here in small quantities, and so are the nutrient elements of plants.

The eluvial horizon (e) is lightly colored (like ash), as considerable quantities of iron and aluminum compounds (such as their fulvates) were washed from here to the underlying horizon. The content of nutrient elements here is low, as well as the content of clay particles. The horizon has an acid reaction and is unsaturated with exchangeable bases. It is practically devoid of structure but may have some platy “cleavage”.

Part of the substances washed from the H_0′ HE, and E horizons are retained in the illuvial horizon (I). This horizon is rich in clay particles, sesquioxides of iron and aluminum, and other compounds. It may be very compacted and of reddish-brown color.

Thus, the development of podzolic process is characterized by a set of the following specific features:

1) biogenic factor is responsible for the differentiation of soil profile into eluvial and illuvial layers under conditions of acid hydrolysis;

2) formation on the surface of the soil of a layer of forest litter consisting of the residues of forest vegetation at various stages of decomposition;

3) transition of the products of litter decomposition into soil profile occurs by the way of leaching and by the lateral subsurface runoff;

4) a considerable number of the litter decomposition products are carried away by the water of surface runoff during snowmelt when the soil remains frozen or filled with water to full capacity.

A pedogenic process of sod formation was subjected to detailed studies by the academicians V.R.Williams and I.V.Tyurin as well as a number of less noted scientists. Grassland vegetation is supposed to be the essential agent of this process. The soils formed are endowed with a well developed surface horizon of humus accumulation. Sod formation in the Polissya zone is characterized by the following features:

1) intensive consumption of biogenic elements and their annual return to the soil;
2) considerable annual accumulation of biomass and its dying out and return to the soil;
3) considerable weight of roots relative to the weight of above-ground plant parts and considerable return of died out root biomass directly to the soil;
4) mostly bacterial decomposition of organic residues and intensive accumulation of soil humus;
5) humatic type of humus and accumulation of humatic acids bonded with calcium

In Polissya zone the sod-forming process occurs on forest glades or on the areas with thinned out forest. It is especially intensive in deciduous forests with well developed grass cover. Eluvial horizon (E) may become richer in SOM and exchangeable bases $(Ca^{2+} + Mg^{2+})$ and gradually turn into a humuso-eluvial horizon (HE).

But even prolonged development of grasses in the forest is unable to cause the accumulation of considerable amount of humus and nutrient elements in soddy-podzolic soil. Plant residues on a poor soil are relatively poor in nutrient elements and bases. Their nitrogen content is low. Water infiltration through the soil brings about additional impoverishment of its horizons in ashy elements. The presence of fungal microflora impedes the mineralization of residues, and fulvic acids dominate in soil humus.

Relative intensity of podzolic and sod-forming processes modified by hydrologic, lithologic and other conditions favors the formation of soddy, soddy-podzolic, podzolic-soddy, and soddy-meadow soils.

Periodic or continuous excessive wetness of Polissian soils strengthens the processes of SOM accumulation and may cause the accumulation of peat. The soils of swampy-podzolic type may have surface horizons H_t or $T_{h.}$

Gleization and peat formation are two stages of the swampy process of soil formation**.** Peaty soils are known in the wored as histosols. They are organic soils, water-saturated for most of the year unless they have been drained. They may contain organic soil materials to a significant depth. Swampy process of soil formation in the Polissya zone may lead to the formation of peat soils of which the lowland peats are dominating.

25.3. Essential Soil Types, Their Diagnostics and Management

The dominant soil types in the Polissya zone include soddy-podzolic, podzolic-soddy, soddy, meadow, and swamp soils. To a much lesser extent there also may be found grey forest, podzolized, chernozemic-meadow, and even chernozemic soils.

25.3.1. Soddy-Podzolic Soils

About 2,489 thousand hectares of soddy-podzolic soils are used in agriculture. They are represented by automorphic (34.1% of farmland area), hydromorphic from the surface 97.1%), and deeply hydromorphic (58.7%) types. The soils eroded by water occupy 3.1% of farmland area, eroded by wind – 0.6%, and drained – 13.7% (M.I. Polupan, et al., 1988). Classification of soddy-podzolic soils of the Ukrainian Polissya is given in Table 58.

Table 58. Classification of Soddy-Podzolic Soils (Ukrainian Research Institute of Soil Science and Agrochemistry, 1988)

Type	Subtype	Genus	Species
Soddy-podzolic	Soddy-podzolic	Modal. Gleyed from the underlying rock upward	By the extent of podzolic process development, gleying, erodedness, and depth of underlying by other rocks

Type	Subtype	Genus	Species
Soddy-podzolic gleyed from the surface	Soddy-podzolic gleyed from the surface	Modal, acid	By the extent of gleying and podzolic process development
Soddy-podzolic gleyed	Soddy-podzolic gleyed Podzolic-soddy-gleyed	Modal, saline	By the extent of gleying

The most widespread genuses of soddy podzolic soils are the following:

1) modal (ordinary) with features that are most typical in a given subtype. The word *modal* is not included into the term naming such a soil;

2) gleyed from the underlying rock upward with a characteristic gleyed layer on the contact of parent material and underlying rock;

3) eroded, which differ from the modal genus by complete or partial loss of upper horizons as a result of water erosion;

4) saline, which form in areas with high watertable of saline ground water. They are found among gleyed soddy-podzolic soils.

The species of soddy-podzolic soils are identified by the extent of podzolic process development:

1) **weakly podzolized soils** with a well developed HE horizon 20-25 cm deep, thicker than the E horizon underlying it (HE>E); eluvial horizon may be represented by separate spots or layers up to 5 cm thick;

2) **moderately podzolized soils** with HE horizon 18-20 cm thick, roughly equal in thickness to E horizon (HE = E). Eluvial horizon is unbroken, with lower boundary more than 30 cm deep.

By the character of soil wetness the following species are identified:

1) **gleyed from the surface** with the upper part of soil profile subject to reduction processes causing gleying. Excessive wetness is seasonal, with precipitation waters ponding on the surface;

2) **deeply gleyed** with gleying features in the lower portion of soil profile, excessive wetness being caused by ground waters.

By the extent of reduction processes (gleying) development the following species of soddy-podzolic soils are identified:

1) **slightly gleyed from the surface** with the signs of reduction processes in humuso-eluvial and eluvial horizons (HE_{gl}, E_{gl}) in the form of separate rusty spots;
2) **gleyed from the surface,** with the unbroken signs of reduction processes in humuso-eluvial, eluvial and the upper portion of illuvial horizons (HE_{gl}, E_{gl}, and I_{gl});
3) **slightly gleyed from ground water,** with ground water table 180-250 cm deep, gleyed parent material and the lower portion of illuvial horizon ($I_{(gl)}$, P_{gl}), signs of gleying being in the form of rusty spots;
4) **gleyed from ground water,** with ground water table 120-180 cm deep, gleyed parent material and the entire illuvial horizon (I_{gl}, P_{gl}), the signs of gleying being of unbroken character, all over the horizons; and
5) **very gleyed,** with ground water table 60-120 cm deep, and distinct signs of gleying all over the profile.

Soddy-podzolic soil may be eroded to different extent with the identification of the following soil species:

1) **slightly eroded soils** with up to one-half of HE horizon thickness and up to 20% of initial amount of SOM lost. Plowing involves the upper part of the Eh horizon, plow layer being lighter in color and with a brownish tint as compared with uneroded soil. Such soil species may be encountered on the slopes up to 3°;
2) **moderately eroded soils** with lost over a half of HE horizon and from 20 to 40% of the initial amount of SOM. Eh horizon is involved into plow layer, as well as the upper portion of EI horizon. Soil surface is brown and spotty. Such soils occur on the slopes 3 to 5° steep,
3) **severely eroded soils** with lost HE, E, and partly I horizons as well as 40-60% of the initial amount of SOM. The middle or even lower part of EI horizon being involved in the plow layer, the soil surface is light brown with large soil blocks covering it. Such soil species may be found on the slopes over 5° steep.

By the depth of underlying by another rock two soil species are identified:

1) with shallow underlying (0.5-1.0 m);
2) with deep underlying (1.0-2.0 m).

The extent of soil salinity causes the identification of the following species:

1) **deeply saline** with soluble sodium salts (0.1% and more) appearing at the depth of gleying;

2) **saline (solonchaky)** with the soluble salts in HE horizons or in the upper portion of E horizon.

Soddy-podzolic soils may be represented by lithologic series of moraine, glacio-fluvial, lacustrine, and alluvial soils, which may be underlain by sands, solid rocks, chalky marl, etc.

There are virgin, reclaimed, cultured, irrigated, and drained variants of soddy-podzolic soils. The soils' variety is determined by their texture, according to which are sandy, finely sandy, loamy-sandy, as well as light, medium, and heavy loams.

25.3.1.1. Soddy-Podzolic Soils Morphological Features

Soddy-podzolic soils proper are zonal Polissian soils. They cover sandy and sandy-moraine plains. Less widespread they are on moraine hills and ridges. Occaicionally they may be found on the ancient alluvial sands of the "borovi" (*bor* means *pine forest*) terraces of the river valleys even outside the Polissyan zone. These are the soils of the coniferous and mixed forests though they occur are found in deciduous forests too. Soddy-podzolic soils have a profile distinctly differentiated into eluvial and illuvial parts. Humus content is low or very low: 0.6-1% in sandy, and up to 1.5-2% in loamy varieties. Their humus is crude with some charred residues in virgin soil variants. Humus type is fulvatic $(C_{h.a.}/C_{f.a.}=0.3-0.8)$. It becomes more humatic as the texture becomes heavier. Soil adsorbing complex (SAC) is unsaturated with bases and soil solution has an acid reaction.

By the relative intensity of podzolic and sod-forming processes, soddy-podzolic soils may be weakly, moderately, and, rarely, strongly podzolic. Soddy-moderately podzolic soils are most widespread. Their profile is distinctly differentiated into HE, E, and I horizons.

Humusoeluvial (HE) horizon in virgin soils is 18-20 cm deep. In cultivated soils it coincides with a plow layer. It is light grey in color, powdery with blocks, structureless with a characteristic platy "cleavage". Quartz sand grains washed from clay and humus coatings attach a characteristic ashy-grey and even whitish color to the entire horizon. Some charred organic residues can be found in the horizon, especially in virgin soil variants. The transition to the lower (E) horizon is usually very distinct (sharp).

Eluvial (E) horizon is unbroken and deep in virgin variants; in plowland soils it is shortened, having been involved into a plow layer. It is the lightest in color horizon in the profile, being whitish grey. Sometimes it is possible to divide it into two subhorizons, $E_{(h)}$ and E_i, the upper and the lower, respectively. **E** horizon is usually structureless (massive) with a noticeable platyness, or even scaliness. It is rich in amorphous silica, especially in varieties heavier than sand in texture. The transition to the lower horizon **(I)** is distinct.

Illuvial (I) horizon usually has some spots of eluviated (washed from ironous and clay coatings) sand in its upper portion (IE, Ie) but deeper it becomes compacted, of reddish-brown or rusty-brown color and nutty-prismatic in structure (I proper). Its lower portion (IP) is less hard (compacted) and rich in lenses and layers of ironous sand of yellow color.

Soddy-weakly-podzolic soils have their E horizon in the broken form of whitish spots and layers about 5 cm thick of sand particles denuded from their ironous (clay) coatings.

Soddy-strongly-podzolic soils are found on a small area and are therefore not identified as mapping units. Their eluvial horizon is over 20 cm thick, OM content – very low, and reaction – very acid.

Soddy-podzolic "secondarily saturated" soils are found on the areas with not too deep (up to 2 m) underlying with chalky marl in western and eastern "extremities" of the zone (Lviv, Volyn, Rivne, Sumy, and Chernigiv regions). Proximity of carbonatic rocks imparts neutral or close to neutral reaction to these soils. Their absorbing complex is more saturated with bases than that of modal soils, and their differentiation into elavium and illuvium is less distinct, especially in the profiles with shallow (0.5-1 m) underlying by chalky marl or limestone.

Soddy-podzolic gleyed soils are widespread on poorly drained watersheds, in the depressions of sandy plains and on the "borovi" terraces, covered with pine forests, pines being least hard of the plants to grow on light-textured and poor soils. Temporary or periodic ponding are waterlogging cause reduction processes (gleying) in soil profile. Excessive wetness may be caused by felling and cultivation of lands for crop productivity after felling.

Morphological features of glaying are applied on the genetic horizons of a soddy-podzolic soil. By the character of excessive wetness, the soils may be gleyed from the surface or gleyed from ground water. Gleying from the surface is characteristic of the soils formed in moraines or underlain by moraines. They have the signs of gleying in the HE horizon (rusty spots, dull grey color, etc). It is customary to differentiate between the soils gleyed from the surface and slightly gleyed from the surface. The former have the features of reduction processes not only in the HE but also in the E, and the upper portion of I horizon. The lower part of the illuvial horizon and the parent material are not gleyed or may show only some scanty rusty spots. Soils slightly gleyed from the ground water show the signs of gleying in parent material and the lower portion of the illuvial horizon. Soils gleyed (proper) from the ground water the entire illuvial horizon intensively gleyed. Very gleyed soils have the entire profile intensively gleyed and ground water table at depth of 30-120 cm from the surface.

Podzolic-soddy soils have a profile like that of soddy-podzolic soils, distinctly differentiated into the horizons of eluvium and illuvium, but their humuso-eluvial (HE) horizon is much thicker (30-40 cm) and contains up to 2-3% of SOM with more pronounced humatic portion (higher $C_{h.a.}/C_{f.a.}$ ratio) compared to soddy-podzolic soils. Illuvial horizon may also be noticeably enriched with humus. Such soils are more widespread on the left-bank Polissya, especially its lowland province. Very often such soils are gleyed. Ground water with sufficient carbonate salinity may cause the development of saline podzolic-soddy soils.

25.3.1.2. Soddy-Podzolic Soils Composition and Properties

By texture the soddy-podzolic soils are distributed in the following way: loamy sands – 48.2%, coarse and fine sands – 11.7%. Soil profile is differentiated by texture, for example the content of physical clay and clay proper in soil horizon. Eluvial horizon is relatively poorer and illuvial-relatively richer in clay (Table 59).

Mineral part of the soil makes up over 96% by weight. It is composed mainly of silicates, alumino-and ferrosilicates, oxides, hydroxides, salts, and organic-mineral compounds. Silicates and quartz are the main components of the soils mineral part. HE and E horizons are the richest in silica content. I horizon is always less rich in it. Accumulation of silica in the eluvial part of soil profile is explained by washing off of aluminum and iron oxides and hydroxides from eluvial to illuvial portion of the profile.

As can be seen from the table below, there is a relative accumulation of R_2O_3 (Al_2O_3 and Fe_2O_3) in the illuvial horizons, whereas calcium and magnesium oxides accumulate in HE and I horizons. Eluvial horizons are poor in these components which proves their biogenic origin.

Table 59. Chemical Composition and Content of Physical (<0.01 mm) and Proper (<0.001 mm) Clay in Soddy-Podzolic Soils (B.S. Nosko and S.P. Latyshev, 1994)

Indices	Soddy-Podzolic Soil (loamy sand)		Gleyed Soddy-Podzolic Soil (sand)			
	Horizons					
	HE	$E_{(h)}$	P_i	HE	E_{gl}	If_{gl}
Layer, cm						
Content of Particles, (%)						
<0.01 mm	14.9	13.3	19.0	9.06	5.16	14.4
<0.001 mm	8.32	5.87	14.2	3.61	6.34	14.4
Oxide Content, (%)						
SiO_2	92.1	92.8	87.8	91.3	90.0	85.9
Fe_2O_3	0.80	0.74	1.69	0.58	0.82	1.55
Al_2O_3	3.25	3.66	5.46	5.80	5.83	8.44
CaO	0.72	0.60	1.15	0.54	0.19	0.20
MgO	0.22	0.22	0.55	0.26	0.20	0.37
$SiO_2 : R_2O_3$	41.6	37.8	22.5	24.9	24.0	15.5

Soddy-podzolic soils are poor in humus (< 2%), which is concentrated in humuso-eluvial horizons (Table 60).

Fulvic acids in soil humus dominate over humic acids. Such composition of humus explains its acid character and solubility.

Cation exchange capacity of soddy-podzolic soils is low (Table 16.4), and so is its saturation with exchangeable Ca^{2+} and Mg^{2+}. Soil reaction is acid and buffer capacity – low.

Table 60. Content and Composition of Humus in Soddy-Podzolic Soils (Atlas of Soils…, 1979)

Horizon and Depth,cm	Total Humus Content,%	$C_{organic}$, %	$C_{h.a.}$	$C_{f.a.}$	$C_{residual}$	$C_{h.a.}/C_{f.a.}$
			% of C organic			
Soddy-weakly-podzolic sandy soil in glacial outwash						
HE$_{0-10}$	0.90	0.52	20.6	41.6	37.8	0.50
Soddy-moderately-podzolic light loam soil in glacial outwash						
HE$_{0-10}$	1.17	0.67	11.9	29.8	58.2	0.39
E$_{30-40}$	0.35	0.20	25.0	45.0	20.0	0.56
I$_{e\ 40-50}$	0.30	0.17	28.5	35.1	40.6	0.65
Soddy-moderately-podzolic light loam soil in moraine						
HE$_{0-10}$	1.48	0.82	20.2	37.4	42.4	0.54

Table 61. Physic-Chemical Characteristics of Soddy-Podzolic Soils (B.S. Nosko and S.P. Latyshev, 1994)

Indices	Soddy-Podzolic Soil (loamy sand)			Gleyed Soddy-Podzolic Soil (sand)		
	Horizons					
	HE	E$_{(h)}$	P$_{i}$	HE	E$_{gl}$	I$_{gl}$
Layer, cm	0-15	23-33	130-140	0-20	25-35	50-60
pH$_{H2O}$	6.3	6.6	6.2	6.2	6.9	6.8
pH$_{KCl}$	5.5	5.6	5.5	5.5	5.6	5.4
Exchangeable cations, meq. per 100 g of soil:						

Ca^{2+}	2.40	1.50	7.14	1.01	0.89	3.38
Mg^{2+}	0.92	0.84	1.13	0.37	0.28	0.64
Na^{+}	0.14	0.19	0.37	0.04	0.06	0.11
K^{+}	0.18	0.12	0.21	0.06	0.04	0.09
Sum (S)	3.64	2.75	8.85	1.48	1.27	4.22
Hydrolytic acidity, meq. per 100 g of soil:	2.70	1.10	0.98	2.90	1.70	1.90
CEC, meq. per 100 g of soil:	6.34	3.85	9.83	4.38	2.97	6.12
Percentage base saturation, %	58	71	80	41	59	74

The lowest CEC is characteristic of E horizons. It increases in I horizons as they are richer in clay and colloids.

Low organic matter content causes low contents of total and available forms of nitrogen and phosphorus (Table 62). The bulk part of nitrogen is contained in organic matter and becomes available to plants only after its mineralization. Organic manures and mineral nitrogen fertilizers are very effective on soddy-podzolic soils. Total potassium content depends on soil texture and mineralogical composition. It fluctuates about 1.1-1.4%, available potassium in A.V. Kirsanov's extract is within 77-140 mg K_2O per 1000 g of soil.

Soddy-podzolic soils are very poor in micronutrient elements: Co:1-2 ppm, Mn: 70-95 ppm, Zn:20-30 ppm, and B: 3-4 ppm.

Physical and related to water movement and availability properties of soddy-podzolic soils are shown in Table 16.6.

Infiltration coefficient of I horizons are lower than those of HE horizons which can cause the excessive wetness of the former and development of reduction processes. Soil aeration can be unstable especially in gleyed soils. Plants growth depends very much on the frequency and amount of rains.

Table 62. Agrochemical Characteristics of Soddy-Podzolic Soils (B.S. Nosko and S.P. Latyshev, 1994)

Indices	Soddy-Podzolic Soil (loamy sand)			Gleyed Soddy-Podzolic Soil (sand)		
	Horizons					
	HE	$E_{(h)}$	P_i	HE	E_{gl}	I_{gl}
Layer, cm	0-15	23-33	130-140	0-20	25-35	50-60
Total Content, % Humus	0.87	0.10	0.01	0.95	0.70	0.07
N	0.05	0.04	Not determined	0.07	0.05	0.02
P_2O_5	0.07	0.05	0.03	0.05	0.03	0.01
K_2O	1.11	0.68	0.66	1.35	1.35	1.39
Available nutrient content, mg per 1 kg of soil:						
P_2O_5	69.5	Not determined	>> – >>	150	Not determined	>> – >>
K_2O	77.7		>> – >>	140		>> – >>

The nature and properties of soddy-podzolic soils have the following specific features:

1) distinct differentiation of the profile into eluvial and illuvial part with the layer of forest litter on the surface;
2) impoverishment of eluvial horizon in respect of clay particles and sesquioxides with their accumulation in the illuvial horizon;
3) relative enrichment of the eluvial horizon with SiO_2;
4) very low content of SOM (humus) in HE and E horizons;
5) the dominance of “aggressive” fulvic acids in soil humus and of brown humic acids over grey ones;
6) the dominance of fungi among the decomposers of organic residues;

7) high values of all types of soil acidity (active, exchangeable, and hydrolytic).
8) low CEC and PBS;
9) high content of soluble Al^{3+}, which increases exchangeable acidity and soil toxicity;
10) low supply of available nutrients;
11) unfavorable physical conditions for plant growth and development.

Table 63. Physical and Soil Water Related Properties Of Soddy-Podzolic Soils (B.S. Nosko and S.P. Latyshev, 1994)

Indices	Soddy-Podzolic Soil (loamy sand)			Gleyed Soddy-Podzolic Soil (sand)		
	Horizons					
	HE	$E_{(h)}$	P_i	HE	E_{gl}	I_{gl}
Layer, cm	0-15	23-33	130-140	0-20	25-35	50-60
Particle Density, g/cm^3	2.63	2.64	2.68	2.63	2.65	2.67
Bulk Density, g/cm^3	1.48	1.61	1.67	1.34	1.42	1.50
Total porosity, % of volume	43.8	39.0	37.6	49.1	46.6	43.8
Maximum hygroscopic moisture, %	1.50	1.50	1.80	1.00	0.80	1.20
PWP,%	2.25	2.10	2.70	1.6	1.20	1.80
FC, %	14.3	12.5	9.70	15.0	18.0	11.0
Available water at FC, %	12.0	10.4	7.0	13.5	11.8	9.2

Potential productivity of soddy-podzolic soils, according to A.I. Siry, is low and very low. Soddy cryptopodzolic soils which are light-textured and with low water-holding capacity have a potential productivity (bonitet) within 17-22 points on 100-point scale.

Soddy-moderately-podzolic soils which are sandy loams, loams, and loamy sands, are potentially most productive (up to 30 points). Though their humus content is low and very low, the amount of available water at field capacity (FC) may reach 160-190 mm in the upper 1-meter layer of soil.

25.3.1.3. Agricultural Use and Practices of Soddy-Podzolic Soils Improvement

Soddy-podzolic soils have the following main causes for low productivity: light texture, excessive wetness, low content of soil humus and macro- and microelements of plant nutrition, and excessive soil acidity. Unfavorable physical properties may also be added to this list.

Light-textured (sandy) weakly and moderately podzolic soils are covered mainly by the forests. Only about 30% of their total area belong to plowland. It is possible to grow winter rye, buckwheat, potatoes, barley, oats, and lupine on such soils. Clover and alfalfa may show only a very low crip-productive power on them. On soils underlain by moraines or loams it can be recommended to grow winter wheat and flax provided that it is possible to ensure a corresponding system of crop fertilizing. Light-textured soils are also no good for orchards. Soil tillage should be performed to varying depth, including subsoiling.

Sandy soddy-podzolic soils are usually plowed to the depth of 18-20 cm in the fields w3ith intertilled crops which are fertilized with organic manures. Farm manure and composts are the essential manures which should be applied at rates of 18-20 mt per hectare of a crop rotation per year.

Regular lime application is also a necessity. Lime should be applied in the fields with the most unresponsive to liming crops in rates from 1.0 to 3.5 mt/ha of $CaCO_3$ every 5-6 years. Sandy soddy-podzolic soils which are loamy sands and loams in texture are most valuable and effectively used soils in the Polissya zone.

About 1,155 thousand hectares of such soils are a plowland (46% of the total area of Polissian plowland). Slightly gleyed species are also included here. Such soils are potentially more productive than sandy varieties. They contain more humus, available water, and nutrients. All Polissian crops can be grown on such soils, including winter wheat, flax, corn, beetroots, clover, etc. Loamy sands may be good for orchards (sweet cherries). Soddy-podzolic loamy soils are also quite suitable for orchards. Primary soil tillage should be performed to varying depth. Soils with deep HE horizon may be plowed to the depth of 28-25 cm, for row crops, and to a lesser depth (18-20 cm), for other crops. Moldboard plowing can be combined with disking and V-blade cultivation. Farm manure should be applied at rates 15-20 mt per hectar of a crop rotation per year. Green manure crops (lupins) may also be useful.

Perennial grasses in crop rotations are a necessity to preserve the level of SOM content and increase the content of available nitrogen, phosphorus, potassium, calcium, magnesium, etc. Perennial grasses also improve physical and physico-chemical properties of soils and intensify microbiological processes in them. Lime should be applied at greater rates than on sandy soils. Usually they amount to 0.75 of the complete rate computed by hydrolytic acidity. Lime is applied once in 8-10 years in crop rotations with lupine, flax, and potatoes, immediately before the preplanting cultivation. Lime can be applied before plowing prior to the planting of crops preceding forage root crops, clover, or corn.

Agricultural crops on soddy-podzolic soils need nitrogen and potassium, and, to a lesser extent, – phosphorus. Subsoiling is well reacted on by clover, forage beets, and forage beans; also favorably responsive to it are corn, timothy grass, and potatoes.

Soddy-podzolic gleyed soils have low potential productivity due to unfavorable aeration, high level of soil acidity and toxicities it involves and not sufficient space for the development of roots (hardness of illuvial horizons). But such soils are satisfactory for the growth of hemp, small grain crops sown in late spring, and vegetables. Loamy sand soils are good for forage crops and perennial grasses. But they are no good for potatoes and fruit trees. Slitting and mole drainage may prove effective on gleyed soils. Drainage is an expensive practice and must be avoided, if possible. Subsoiling with loosening of a subplow layer and planting on the ridges make another way out of a complicated situation. Lime should be applied at rates of 3 to 7 mt/ha in combination with phosphorite meal which favors the involvement of soluble aluminum into complex compounds so that it looses its toxicity to plants. Organic manure rates should be lower than on nongleyed species. Lime (0.5-0.75 of a complete rate computed by hydrolytic acidity index) should be broadcasted on the surface and incorporated into the soil by heavy disks.

25.3.1. Soddy Soils (Sod Soils)

Soddy Soils (Sod Soils) belong to intrazonal soil types, but are most widespread in the Polissya zone. They occupy 1,706 thousand hectares (about 20% of the zone's area) in Zhitomir, Chernigiv, Kyiv, Volyn, and Rivne regions. The soils there were formed under grasses by the sod-forming process of soil formation the essential feature of which is the accumulation of humus in the upper horizons of soil profile. Soddy soils form in various hydrothermal conditions. They, therefore, are divided into four separate soil types: 1) soddy-carbonatic gleyed; 2) soddy skeletal; 3) pine forest soddy, and 4) soddy gleyed.

25.3.2.1. Classification of Soddy Soils

Soddy-carbonatic soils were formed in carbonatic parent materials: limestones, chalk, chalk marl, loams and clays with limestone concretions. The classification of such soils is given in Table 64.

Table 64. Classification of Soddy-Carbonatic Soils (Research Institute of Soil Science and Agrochemistry, 1988)

Type	Subtype	Genus	Species
Soddy-Carbonatic	Modal, leached, podzolized	Modal, eroded, washed-on	**By the depth of loose (weathered) layer:** slightly developed short-profiled
Soddy-Carbonatic	Modal, gleyed, leached gleyed, podzolized gleyed	Modal	**By the content of SOM:** slightly humusified with low humus content **By the extent of soil erosion:** slightly eroded moderately eroded severely eroded **By the extent of gleying:** slightly gleyed (moderately) gleyed **By stoniness:** not skeletal slightly skeletal moderately skeletal very skeletal

Leached subtypes of soils effervesce with 10% HCl not from the surface but from the transitional horizon. Podzolized soils effervesce only in parent materials and have elluvial-illuvial redistribution of colloids in the profile. All modal soils effervesce from the very surface and the word *modal* is not used in the term denoting this type of the soil.

Slightly developed soddy-carbonatic soils have the surface loose horizon only up to 25 cm deep, **short-profiled:** 25-45 cm deep, and modal: over 45 cm deep. **Slightly humisified soil** species have humus content in the surface horizon only up to 3%. Soil species with low humus content have it within 3-6%. **Slightly eroded soils** lost up to a half of the H_k horizon (up to 20% of the initial amount of humus). Plow layer may contain limestonic skeleton. As a result of erosion, **moderately eroded** species have the entire H_k horizon lost and severely eroded ones – the entire "humusified" profile: (H_k + HP_k) lost. Soil surface is as good as white with numerous limestone fragments.

Slightly gleyed soils have the signs of reduction processes in parent materials, whereas gleyed ones have them all over the profile.

Slightly stony (or skeletal) species have up to 10-30% of rubble or other skeletal particles on the surface of 10 square meters; moderately stone: 30-50%, and very stony: over 50%.

Soddy skeletal soils are divided into subtypes of modal and podzolized soils and the genus of modal soils. Their species are identified by the depth of the loose upper later, by the extent of stonyness (skeletality) and by the content of rubble,

Pine forest ("borovi") soddy soils are encountered on the "borovi terraces" of the river valleys in all soil-climatic zones of Ukraine, being intrazonal. After felling of the forests, such soils become very susceptible to wind erosion. Their texture is so light that only thinned out pine forests can hardly grow on them. Humus content is very low. Soil profile differentiation is very weak. Their water, air, nutrient and thermal regimes are unfavorable for field's crops and other plants. Borovi soddy soils are classified into sub-types of borovi soddy proper, borovi soddy podzolized and chernozem-like soddy soils (Table 65). There are lithologic series of ancient alluvial and ancient alluvial with buried soils units of soil cover.

Soddy gleyed soils lie on the lower elements of topography (within flood plains and flood plain terraces, on the terraces over flood plains, in the watershed depressions, and on the peripheral parts of swamps). They form in alluvial, glaciofluvial, and loamy deposits under meadow, meadow-swampy, and forest vegetation with excessive wetness caused by continuous waterlogging and periodic ponding.

Classification of gleyed soddy soils is given in Table 66.

Table 65. Classification of Borovi (Pine Forest) Soddy Soils (Ukrainian Research Institute of Soil Science and Agrochemistry, 1988)

Subtype	Genus	Species
Soddy borovi	Modal, flown out, ironous, and pseudofibrous	**By the depth of humus layer (H+Hp):** slightly developed (< 25 cm) short profiled (25-45 cm)
Soddy borovi podzolized	Modal	**By the depth of buried soil location:** not deeply buried (0.5-1.0 m) deeply buried (1.0-2.0 m)
Chernozem like soddy	Modal	**By the depth of humus layer:** not deep (45-65 cm) moderately deep (85-120 cm) deep (over 120 cm)

Table 66. Classification of Gleyed Soddy Soils (Ukrainian Research Institute of Soil Science and Agrochemistry, 1988)

Subtype	Genus	Species
Soddy gleyed Soddy podzolized gleyed soddy solodized gleyed soddy gleyed saline soddy gleyed sodic	Modal, carbonatic, marly-carbonatic, saline, washed-out, ortzandic, and ortsteinic	By the extent of gleying, podzolization, sodicity, salinity, and solodicity

The word “modal” is not stated in the soil’s name. The most widespread species of gleyed soddy soils are 1) gleyed, experiencing the effects of high water table and waterlogging and having the signs of gleying in all horizons, and 2) slightly gleyed with signs of gleying in the lower portion of soil profile, formed with changing level of water table and better drainage.

Podzolized gleyed soddy soils show a characteristic redistribution of colloids caused by the effects of forest vegetation.

25.3.2.2. Soddy Soils Morphological Features

Soddy carbonatic soils are also known as *Rendzinas* (*Leptosols Rendzic* by the FAO nomenclature). Such soils are found in the “western extremity” of the Polissya zone where they formed in the deposits of chalky marl. Sometimes soddy-carbonatic soils form in glaciofluvial sands and loamy sands underlain by chalky marl from the depth of not more than 50 cm. Good saturation with bases impedes the development of podzolic process and, in most cases, it does not develop at all. The soil reaction is neutral to slightly alkaline. One of the diagnostic features is relatively high SOM content (2.5-4.0%). Soil profile is composed of the horizons: Hk + HPk + PK. In well developed soil species Hk + HPk = 60-100 cm and even thicker, soil productivity greatly depends on the depth of loose (withered) upper layer as that determines the space for the growth of roots. Soddy carbonatic gleyed soils contain more humus (4-5%) and are better supplied with moisture than their non-gleyed counterparts.

Soddy skeletal (lithogenic) soils are found in the area of the Ukrainian crystalline shield where crystalline rocks (granites, gabbro, and diorites) reach the surface of the earth, or are covered by the mantle of glaciofluvial sand and loamy sands not thicker than 50 cm.

The soils are undeveloped. Their suitability for tillage and use for crop production is determined by the space for roots. Short-profiled species have the upper loose horizon 25-42 cm deep, and modal – over 45 cm. The soil profile may be given as $H_g + HP_g + P_g$. Undeveloped species do not have any transitive horizons.

Pine forest ironous soddy soils are found mainly on sandy (borovi) terraces with ancient alluvial sands in the Polissian and Forest-Steppe zones of Ukraine. Grasses in pine forests are weak and scanty. Soil profile may be composed of the horizons: $H_0 + H_{(e)} + I_f + P$. Morphological features of podzolic process are very indistinct. I_f horizon is called ironous as it contains hard rusty-colored layers.

Pine forest ironous-illuvial species have a profile: $H_e + P_e + P_i$. Pine forest pseudofibrous soddy soils contain the horizons: $H_{(e)} + P_e + P_f + P$. Pseudofibres are 1-2 cm thick, of brownish red color and noticeable hardness.

Pine forest soddy soils proper are found on the sand terraces of river valleys in various zones. Sometimes they occur on sand ridges near the river beds. Their profile is of accumulative type: $H + P_h + P$ or $H + HP + P_h + P$. The soils form not only in alluvial sand but in the sands of glacial outwash, too.

Soddy gleyed soils occur in lowered areas within watersheds or terrace plains and occasionally on sandy ridges within the swamps. Their profile may be described as $H + HP_{gl} + P_{Gl}$. SOM content depends on texture and may be within 1-5%. Slightly gleyed and (properly) gleyed soil species are identified by the extent of gleying. Slightly gleyed soils are the result of periodic waterlogging, gleyed soils – of continuous one. Saline ground waters facilitate the development of salinity-affected or solodized soddy soils. Virgin variants of soddy gleyed soil have a profile described by the sum: $H_d + H_{gl} + P_{hgl} + P_{Gl}$. H_d (sod) is 5-10 cm thick, with a multitude of grass roots.

Soddy-gleyed podzolized soils occur in Прикарпаття on wide terrace plains. On watersheds they are rare. Such soils were formed in ancient alluvial, lacustrine, and loamy outwash deposits. Surface horizons are of a characteristic brownish tint. Soil profiles may be determined by the following sums:

1) soddy podzolized deeply gleyed: $H_e + H_{pi} + I_{gl} + P_{Gl}$;
2) soddy podzolized moderately gleyed: $H_{eh} + H_{igl} + I_{gl} + P_{Gl}$;
3) soddy podzolized severely gleyed: $H_{eGl} + H_{piGl} + I_{Gl} + P_{Gl}$.

25.3.2.3. Soddy Soils Composition and Properties

By their texture, soddy carbonatic soils are mainly light loams (37.4%) and moderate loams (33.8%). Loamy sands occupy 19.6%, heavy loams – 19.6 and sands – 1.2% of the total area (D.I. Kovalishin, 1988).

The texture determines physical and water-related properties of the soils. The bulk density in sandy varieties is within 1.45-1.68 g/cm^3, in light loams – 1.20-1.35, and in heavy loams – 1.18-1.25 g/cm^3. As soil texture becomes heavier, total porosity increases from 44 to 55%, aeration porosity reduces from 34 to 26% and field capacity increases from 40 to 70 mm. But the maximum amount of available water in 0-100 cm layer of soil changes little and remains within 170-175 mm. The content and quality of SOM in soddy soils are given in Table 67.

Table 67. Content and Composition of Humus in Soddy Soils

Horizon, cm	Total humus content, %	Total C_{org} content, %	C h.a.	C f.a.	C residual	C h.a / C f.a.	Type of humus
			% of total C_{org}				
Soddy-carbonatic light loam soil in eluvium of chalk (M.B. Lisovy, 1979)							
$H_{k\ 0\text{-}10}$	3.32	1.93	41.8	38.9	19.3	1.07	fulvate-humatic
$HP_{k\ 21\text{-}30}$	1.50	0.87	37.9	39.1	23.0	0.97	humate-fulvatic
Soddy skeletal loamy sand in glacial outwash (L.V. Kryuchkov, 1974)							
$H_{d\ 2\text{-}12}$	1.83	1.06	26.3	56.5	17.2	0.47	fulvatic
$PH_{\ 25\text{-}35}$	1.13	0.66	25.0	54.3	20.8	0.46	fulvatic

Humus fractions bonded with Ca dominate in the composition of humic and fulvic acids. Percentage base saturation (PBS) is within 30-50 meq. per 100 g of soil. Soil reaction is neutral to slightly alkaline. (pH_{H2O} 7-8).

By chemical composition soddy-carbonatic soils have a weakly differentiated profile. They contain less SiO_2 and more CaO, compared to soddy-podzolic soils. The percentage of CaO increases with depth whereas that of other oxides decreases. The soils are poor in B, Cu, Co, and Zn, but rich in the macroelements of nutrition. At the same lime, their phosphorus is bound in $Ca_3(PO_4)_2[Ca_5(PO_4)_3OH]$.

The essential practices increasing soil productivity are those improving water harvesting and retention and increasing the availability of nutrients. This is achieved with proper soil tillage, and the use of acidifying forms of mineral fertilizers like superphosphate, crude potassium salts, etc. The soils (and crops) need also the application of B, Cu, and Co.

Soils feasible for crop production should have the surface loose layer no less than 25-30 cm deep. Winter wheat, sugar beet, and corn can be grown on well developed soddy-carbonatic soils. But they are not good for potato, and downright bad for lupin, flax, hop, and fruit trees.

Skeletal soddy soils have a high bulk density (1.43-1.66 g/cm^3), unfavorable for agricultural crops. The soils can subside and become compacted after tillage. Their maximum hygroscopic moisture is 1-1.2% and their water penetrability is 300-600 mm per hour which is (excessively high). In summer such soils may become too dry. Humus content may be within 1.1-1.8%, with fulvatic type of humus. The percentage of base saturation is up to 75-80%, soil reaction is neutral (pHH2O 7-7.3), and CEC is within 15-20 meq. per 100 g of soil. Because of skeletality, the soils are less than moderately good for crop production.

Pine forest (borovi) soddy soils, 50 thousand hectares of which are involved in agriculture are mostly coarse and moderately coarse sands. Their fertility is very low, especially the content of microelements like B, Cu, and Zn. Exchangeable calcium and magnesium in the surface horizon reach the levels of 2-6 meq. per 100 g of soil, PBS in plowed variants is within 70-85%, whereas in virgin variants it is only 40-60%. Total nitrogen content is within 0.02-0.03, phosphorus – within 0.07-0.09, and potassium – 0.5-1.0%.

Gleyed soddy soils are more acid (pH_{KCl} 5.5-5.8), with hydrolytic acidity 1.0-2.8 meq. per 100 g of soil and humus content up to 4%.

Potential productivity evaluation shows great soddy soils are generally of low productivity, which, however depends on soil texture and humus content, and may be 40 points (on the 100-point scale) for soddy-carbonatic and only 11 points for pine forest types. In the majority of cases, soddy soils are used as grasslands and pastures; only 20% of their area is a plowland. Pine forest soils are not recommended for agricultural production.

25.3.3. Alluvial Soils

25.3.3.1. Alluvial Soil Genesis

A large group of alluvial (floodplain) soils were formed on floodplain terraces of the river valleys. The wider the river, the wider its floodplain. There are over 600 rivers in the Polissya zone of Ukraine, 20 of which are over 100 km long. Two specific soil forming processes – floodplain and alluvial take place in river valleys.

Floodplain process shows itself in periodic floods on floodplain terraces. The soils here may be flooded for a short (7 days), moderate (7-15 days), long (15-30), and very long (over 30 days) period of time. Flooding water drastically changes soil water and the aeration regime of flooded area, softens the fluctuation of temperature, creates a specific microclimate, determines the character and intensity of microbiological processes and affects the composition of plant associations and their productivity.

The character of a floodplain process is very important for the agricultural use of alluvial soils. Short inundation way be tolerated by most crops, moderate one cannot be tolerated by winter small grain crops, and long flooding period does not allow to grow fruit trees. Very long flooding period can be tolerated only by the hydrophytes.

Alluvial process is the accumulation of alluvium deposited by the river. Every year a layer of silt is deposited on the floodplain which becomes involved in soil formation. Alluvial soils, therefore, continuously grow upward systematically receiving new portions of parent material. Ground water is an unalienable agent of alluvial soil formation. Ground water level changes at the time of floods. Alluvial process enriches the soil with nutrient elements, silt, etc. It also forms the relief of a floodplain and favors the spreading of different plant species supporting in this way the fertility of alluvial soils and determining the character of their use in crop production. According to V.R.Williams, a well developed floodplain contains three parts (zones): 1) that bordering with the river bed; 2) central, and 3) adjacent to the terrace (Fig. 96), which differ from each other by the composition of alluvial deposits, character of landforms, hydrologic conditions, and, as a consequence, by the character of vegetation and soils.

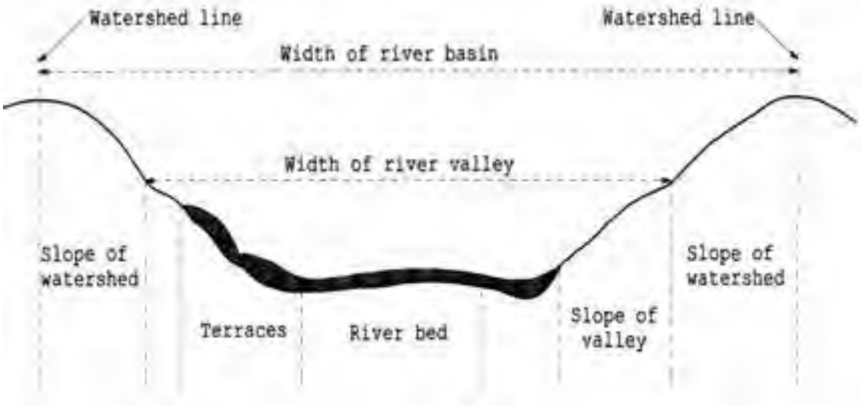


Fig 96. Transsection Across the River Valley (M.P.Tolstoy, 1991).

Floodplain bordering with the riverbed is a narrow band along the riverbed. In small rivers it is 20-50 cm wide, but in large rivers it may be a few kilometers wide. It is composed of coarse material of non-uniform texture, which is the reason for its being called “layered”. Usually it is sandy, its surface is wavy, with sand ridges. Undeveloped soils of low fertility form here.

Central floodplain is the widest portion of a river valley. It is formed of silty and clayey particles. The soils here are layered too. During snowmelt and spring floods the water brings here a non-sorted material. Crude particles form make up sediment first. As the current becomes slower, some finer material subsides forming layered alluvial deposits. Land surface on central floodplain becomes level, with some elevated ridges and manes. The meadows of three types are formed in central floodplain: insufficiently, moderately and excessively wet. The meadows are characterized by perennial grasses and may be used for many years. The former riverbeds are surrounded by bushy vegetation. Hydrologic regime determines the character of soils – from soddy to alluvial swampy.

Floodplain adjacent to the terrace is the fartherest from the riverbed and the lowest part of a floodplain. Deposits here are heavy textured and so are the soils formed in them which may be even clays. Water-loving (hydrophytic) plants with long stems grow here. The soils are meadow swampy and gleyed silty and peaty soils. Ground water is also effective in soil formation.

Alluvial soils may form under conditions of continuous, but not always occurring every year, flooding. The layers of silt accumulate on the surface. Alluvial process of soil formation is not uniformly intensive all over this part of the plainland. So alluvial soils are very different and form a complex soil cover.

Alluvial soils are hydromorphic. They are also intrazonal being found red in different soil-climatic zones of Ukraine. About 632 thousand hectares of such soils are a farmland of which 235 thousand hectares are a plowland.

Alluvial sod soils develop with short-term wetting by spring floods. Ground water with low watertable and capillary fringe which remains below soil profile for a greater part of the year does not significantly effect soil formation. The deposits of the floods are light-textured, poor in organic matter and exchangeable bases.

Alluvial meadow soils develop with excessive wetness from both floods and ground water, with the water table not deeper than 1-2 m. Capillary fringe remains within soil profile. Biogenic accumulation of substances in the upper horizons of soil profile depends to a considerable extent on the substances dissolved in ground water. Yearly alluvial sediments which are heavy-textured and rich in bases and organic compounds play an important role in soil formation,

Alluvial bog soils develop under conditions of long-term excessive wetness caused by the floods, ground water, and precipitation. The soils are characterized by the accumulation of underdecomposed organic residues as well as of substances incoming with ground water and floods. Alluvial swampy soils, by soil reaction, composition, and properties are divided into three groups: acid, saturated with bases (neutral or very slightly acid), and carbonatic (with slightly alkaline reaction and saturated with bases).

25.3.3.2. Classification, Properties, and Use of Alluvial Soils

Classification of alluvial soils is given in Table 68. The soils are divided into three types: alluvial soddy, meadow, and meadow swamp soils.

Table 68. Classification of Alluvial Soils (Ukrainian Research Institute of Soil Science and Agrochemistry, 1988)

Type	Subtype	Genus	Species
Alluvial sod soil	Modal, primitive layered podzolized	Modal, acid, layered, gleyed	By the depth of SOM accumulation, extent of podzolic process development, and gleying
Alluvial meadow soil	Modal, podzolized, sodicity-affected, solodized, brown-earthic	Modal, carbonatic, layered, podzolized from the surface, deeply sodicity-affected, and gleyed	By the depth of SOM accumulation, by the depth of carbonates, by the extent of layering, podzolic process development, affection by sodicity, salinity, and gleying
Alluvial meadow-swampy or bog soil	Modal, sodicity-affected, solodized	Modal, peaty, layered	By the extent of sodicity and solodicity

Alluvial soddy soils are formed mainly under meadow vegetation, but some place they occur under “stepped” bushland vegetation and forests situated close to riverbeds under the impact of short-term flooding. After flooding, the boundary of capillary fringe subsides below the solum (soil profile over parent material). The soils are not excessively wet and show no signs of gleying. Soil profile may be given as $H_d + (H) + P_h + P$: H_d is slightly compacted earthen looking sod, usually of low thickness.

H is the horizon of humus accumulation, cloddy or grainy in structure. Its thickness, depending on the extent of soil development and intensity of alluvial process, may be within 3-20 cm.

Ph is transitive layered horizon. Not always well developed especially in shallow undeveloped soils.

P designates alluvial deposits of varying texture. In floodplain part, bordering with a riverbed, it is light-textured and layered.

The soils are characterized by high infiltration coefficient, good aeration, dominantly downward flow of water, high redox potential, acid reaction ($pH_{H2O} < 6$), and low PBS (10-20%). Humus content is about 1-2%. Humic acids slightly exceed fulvic ones. The nutrient content is varying, depending on soil composition.

Alluvial soddy layered undeveloped (primitive) soils are the youngest in a floodplain. They develop on the ridges and “islands” near the riverbed. Their profile may be summed up as: $P_1h_1 + P_2h_2 + \ldots + P_nh_n$. The soils are, as the name implies, poor in nutrients and slightly saturated with bases. Such soils are used as pastures.

Alluvial soddy layered soils form in crude alluvium near the riverbeds. They are “seated” 3-4 m above the water level and develop under floodplain forests or meadow grasses. Their profile is composed as follows: $H_d + (H) + P_h + P$. The soil reaction varies from slightly acid to neutral (pH_{KCl} 5.5-6.7). The soils are poor in bases and nutrients. They are employed mainly in grasslands.

Alluvial modal soils are situated on the central floodplain, on the elevated landforms. They have a deep horizon of humus accumulation (20-30 cm) of well granulated structure, and have the following system of horizons: $H_d + H + PH + P$. Their surface layer may contain up to 5% of SOM in which fulvic acids dominate over humic ones. Soil reaction is slightly acid or acid (pH_{KCl} 4-6) under forests. CEC is within 10-20 meq. per 100 g of soil. After plowing and a good fertilizing the soils become good for the production of vegetables.

Alluvial soddy podzolized soils are found on the elevated portion of a floodplain, 6-7 m above the river level, and they seldom if ever suffer from floods. The soils are under forests and meadows. They are composed of the horizons: $H_d + H_{(e)} + PH_{(i)} + P$. Loamy soils may have quite acid reaction (pH_{KCl} about 4). Sandy varieties are less acid (pH_{KCl} 4-5). SOM content (3-5%) sharply decreases with depth. Fulvic acids dominate in soil humus. For such soils to be used for crop production it is necessary to use their liming.

Alluvial meadow soils cover central floodplains, and form under meadow grasses, suffering flooding, in sandy-loam or loam parent materials. After the flood subsides, the upper part of capillary fringe still remains within the profile, so the lower horizons show the signs of gleying: H_d + (H) + $HP_{(k)}$ + $Ph_{(k)gl}$. The most characteristic soil features are: optimum (sometimes excessive) moisture of humus horizons, high water-holding capacity, alteration of downward water movement with upward one, seasonally unstable redox potential, and acid reaction ($pH_{H2O} < 6$). The soils may contain toxic amounts of iron (Fe^{2+}) in the lower profile. In H horizon, SOM content may be within 4-6%. There may also be carbonatic and other subtypes. Alluvial meadow swampy soils form near the terraces, in continuous excessive wetness, under "swamp-loving" vegetation in alluvial deposits of loamy texture. Every year the soils are flooded for over 30 days. The character of ground water composition and regime causes the formation of saline, ironous, carbonatic, and other soils. There are two subtypes of soils: alluvial meadow-swampy (H_d + H + HP_{gl} + P_{gl}) and peaty alluvial meadow-swampy (T_h + H_{gl} + Ph_{gl} + P_{gl}).

Meadow soils form under meadow grasses on the plateaus, terraces, ravine bottoms, etc. Ground water plays an active part in their formation. N.B.Vernander came to a conclusion that with ground water table 1-2 m deep, meadow soils form in loamy deposits, with ground water table 1-1.5 m deep – meadow-swampy and with ground water table less than 1 m deep – swampy soils.

Meadow soils are used as farmland (1,206 thousand hectares) and plowland (478 thousand hectares). Polissian meadow soils are mainly short-profiled: H_d + H + HP + Ph_{gl} + P_{gl}. Podzolized meadow soils form on the spots once covered by the forests: H_e + HP_i + Ph_{igl} + P_{gl}. Sands and loamy sands are poor in SOM (1.5-2%), loams are richer in it (3-5%). SOM is of a fulvatic type. Such soils are used as productive grasslands.

25.3.4. Swampy Soils

They are also called bog and peat soils. The area of peaty-swampy soils in Ukraine is 1,202 thousand hectares, of which about 570 thousand hectares are a farmland. By the conditions of water and mineral nutrition, the swamps are divided into upland, transitive, and lowland. The first two occupy a very limited area in Ukraine and are found mainly in Polissya zone: the north of Rivno, the north-west of Zhitomir regions, and in the Carpathians. Their area does not exceed 5% of the total area of swamps.

Lowland peat soils occupy 635 thousand hectares in the Polissya zone.

24.3.4.1. The Origin of Bogs

Bogs were formed in excessive wetness from ground water, or from the surface (precipitation), or both. There are two essential types of bog formation: 1) swamping of dry land depressions and 2) peat formation in the reservoirs of inland water. Any way, bog soils are the soils with aquic moisture regime. Acquic and udic (humid) moisture regimes are the essential features in the bog soil formation process which is composed of two phases: 1) gleying, and 2) the accumulation of peat.

Swamping of dryland. Soil surface or the lower soil horizons are saturated continuously during the year or in some period of time. Water saturation occurs because of the location of the soil in a landscape position that is naturally poorly drained or where an impermeable layer causes saturation. Mineral soils with aquic regime are frequently dull grey, which indicates reducing conditions and low oxygen supply. Alternating reducing and oxidation conditions are associated with a fluctuating water table that produces mottled color. Gleying is followed by the accumulation of peat. Upland oligotrophic swamps are formed when precipitation exceeds evaporation and the soil’s water penetrability is low. Surface runoff and river water cause the formation of transitive or lowland swamps near the foothills of the slopes or in the river valleys. Excessive wetness caused by ground water forms **lowland swamps**.

Lowland swamps, understandably are richer in nutrient elements, both macro- and micro-, than upland swamps. And the evolution of vegetation differs in these two types of swamp formation.

Swamping of dryland, though, is not typical for the territory of Ukraine. Peat accumulation in the inland water reservoirs dominates here.

The lakes and other reservoirs, overgrown by vegetation, are becoming more and more shallow, because of the accumulation of organic debris. They encourage the growth of many plants such as pondweed, cattails, sedges, reeds, and other grasses, as well as mosses and shrubs. These plants in numberless generations thrive, die, and sink down, to be covered by water which shuts out the air, and prohibits rapid oxidation. The decay that does go on, is largely through the agency of fungi, anaerobic bacteria, algae, and certain types of microscopic aquatic animals. As one generation of plants follows another, layer after layer of organic residues is deposited in the swamp or marsh. The profile of an organic deposit is, therefore, characterized by layers differing not only as to their degree of decomposition but also as to the nature of the original peant tissue.

The ecological role of bogs displays itself in two aspects. Swamps and bogs are **geochemical barriers** accumulating certain elements, **mechanical barriers**, which retain fine earth which is washed off the surrounding slopes, and **physico-chemical barriers,** which retain and preserve a number of valuable substances, migrating in biosphere, and transfer them from geologic to biologic cycle due to the accumulating role of swamp vegetation.

25.3.4.2. Classification of Bog Soils

Polissian swampy soils have a deep layer of peat and much more rarely form in floodplains than elsewhere. In the Forest-Steppe and Steppe zones, they are mainly in floodplains, and their peat horizons are less deep, often saline, carbonatic, or sodicity-affected.

Classification of bog soils is given in Table 69. Organogenic bog soils are classified by the depth of peat layer: peatish-gleyed (T < 30 cm), peaty-gleyed (T = 30-50 cm), and peats (T > 50 cm), the latter may be not deep (50-100 cm), moderately deep (100-200 cm), deep (200-400 cm), and extra deep (> 400 cm). Silty-peaty soils have alluvial (T_{al}) or deluvial (T_{dl}) horizons on the surface.

Table 69. Classification of Peat Soils

Subtype	Genus	Subgenus	Species
Type: Upland and Transitive Bog Soils			
Peatish-gleyed (T < 30 cm) Peaty-gleyed (T = 30-50 cm), Peats (T> 50 cm),	Modal, transitive, humusoironous	By the botanical composition of organogenic horizon: ligneous-mossy, grassy-mossy	By the depth of peat layer and the extent of peat decomposition
Type: Swampy Lowland Soils			
Organogenic Soils			

Subtype	Genus	Subgenus	Species
Peatish-gleyed (T < 30 cm) Peaty-gleyed (T = 30-50 cm), Peat soils (T> 50 cm), Silty-peaty soils (T_{al}, T_{dl})	Modal, carbonatic, saline, sodic, ironous, silted	By botanical composition of organogenic horizon: grassy, ligneous-grassy	By the depth of peat layer, by the extent of peat decomposition and humification, ash content, salinity, drainage, etc.
B: Mineral Soils			
Swampy mineral (H_d(T)) Silty-gleyed (H_{dg}l) Humuso-gleyed (H_{gl})	Modal, carbonatic, marly, sodic-saline	>> – >> –>>	By the extent of humus layer, by the content of SOM, by the presence of carbonates, type and extent of salinity

Among the subtypes of swampy lowland mineral soils identified are the soils with $H_{d(t)}$ horizon, silty-gleyed soils with H_{dgl} horizon, and humified gleyed horizon (H_{gl}).

By the extent of peat decomposition identified are peaty soils (peat decomposition up to 25%), and peaty-humusified ones (peat decomposition within 25-45%). In humusified surface horizon the peat soils extent the hand and it is impossible to identify any plant residues in its mass.

Botanical composition is determined only for slightly and moderately decomposed peats. By the extent of ash content there are low ashy (up to 8% of ash), moderately ashy (8-20%), highly ashy (20-50%, silty-peaty (50-80%) and mineral (over 80%) swampy soils.

25.3.4.3. Properties and Use of Bog Soils

Soils of upland and transitive bogs have low ash content (2-6%), high level of acidity (pH_{H2O} 3.5-4.5), and a characteristic convex surface. Their profile is composed of peat horizon (T) not more than 30 cm deep, brown in color, fibrous with mossy sod on the surface (T_d), consisting of undecomposed sphagnum, and gleyed mineral horizon, which is conventionally considered a parent material.

Soils of lowland bogs were formed in lowland poorly drained land forms under native vegetation represented by sedges, reeds, cattails, pondweed, mosses, alder, horse-tail, etc. The ash content of peat is moderate to high. Soil reaction is in the wide range of pH 5-8.5. Nutrient content is considerably higher than in upland peat.

Mineral swampy soils in their virgin variants have a profile with the horizons: $H_{d(t)} + H_{gl(t)} + PH_{Gl} + P_{Gl}$.

Moderately deep peat soils of various ash content in their virgin variant have the following sequence of horizons: $T_0 + T_1 + T_2 + T_3$ and so forth. A drained variant of soil profile has $T_1H + T_2H + T_{3h}$, and so forth. Underlying material may be very gleyed mineral rock (P_{Gl}) or gley (Gl).

Peat soils are potentially productive, especially lowland peats rich in nitrogen, calcium, sometimes even phosphorus, and characterized by neutral to slightly acid reaction. Peat soils, however, are characterized by a number of unfavorable properties limiting their use in agriculture: high water holding capacity, high heat capacity, low bulk density, and low thermal conductivity. Well decomposed peats have a low infiltration rate. Slightly decomposed peats have a low capillary fringe.

Drainage of swamplands does not improve many negative properties of peat. Their density increases and water retention decreases. Carrying capacity of peat soils with moisture 80-85% by weight does not exceed 0.5-1.0 kg/cm^2. Some heavy tractors and machinery should not be used on such soils. Excessive nitrogen in peat soils causes the lodging of small grain crops, especially winter wheat. Moisture unavailable for plants may be not less than 35-40% of full capacity. Peat soil productivity evaluation criteria still need to be identified or improved. Their potential productivity index, according to A.I. Siry, is within 20-35 points on a one-hundred point scale.

Research results and practice of farming on such soils allow to name the following practices for improving swampy soils:

1) drainage, the construction of drainage-watering systems;

2) the roots and stems of natural vegetation must be given time to decay or removed, but by no means burned out;

3) it is preferable to break the peat with heavy plows to a sufficient depth;

4) lime and other calcium compounds should be applied regularly;

5) for a year or two it is advisable to grow such crops as corn, oats, or rye. Then the soil may be used for vegetables production;

6) all sorts of vegetables may be grown on peat soils: celery, lettuce, spinach, onions, potatoes, beets, carrots, asparagus, and cabbage;

7) complete reliance is placed on mineral fertilizers. Phosphorus as well as potassium must be applied in large amounts; copper sulfate and the salts of Mn and Zn should be applied too.

8) Bog ecosystems suffer very much from excessive recreation loadings and other anthropogenic agents. Useful functions of natural swamps and negative effects of drainage dictate a different policy in the use of swampy soils. The best practice to be recommended would be to leave them in their natural state and not to undermine their functioning as "lymphatic nodes of the biosphere".

Questions & Assignments

1. Describe natural features of the Forest Zone.
2. Characterize parent materials of the Forest Zone.
3. Describe podzolic process of soil formation.
4. Speak on the classification of soddy-podzolic soils.
5. Speak on the use of soddy-podzolic soils in agriculture.
6. Characterize soddy soils (their genesis, classification, and morphologic features).
7. Speak on the swamp (bog) soils in Ukraine (their area, genesis, and use).
8. What do you know about the environmental importance of bos?

Chapter 26

Soils of the Forest-Steppe Zone of Ukraine

The Forest-Steppe soil climatic zone in Ukraine extends for 1500 kilometers as an unbroken band from the Carpathian foothills in the west to the state boundary with Russia. The width of this band is within 150-330 kilometers, it stretches from the south-west to the north-east as in this direction the climate becomes more arid and continental. Its northern boundary is natural and quite distinct whereas its transition to the Northern Steppe zone is less clear and goes along the river of Oskol and further along the line: Izyum, Petrivka, Pereshchepino, Kotovka, Tsarichanka, Verkhnyodniprovsk, Olexandria, Adzhamka, Pishchany Brid, Kryve Ozero, Lubashivka, and Krasni Okny.

The zone is distinctly differentiated by the nature into three physico-geographic provinces: Prycarpathian, Right Bank, and Left-Bank. The Prycarpathian soil province (the western Forest-steppe) is most complicated due to its landforms and other natural features. It includes the Volyno-Podilsky upland, Opillya, and Roztochchya (Ternopil, Khmelnytsky and northern Chernigiv regions).

The Right-Bank Forest-Steppe occupies the southern parts of Zhitomir and Kyiv regions, greater parts of Khmelnitsky, Vinnytsya, and Cherkasy regions, and the northernmost districts of Odesa and Kirovograd regions. The Left-Bank Forest-Steppe includes the Prydniprovska lowland along the rivers of Seym, Desna, and Dnipro, enveloping some parts of Sumy, Chernigiv, Kyiv, Cherkasy, and Poltava regions, and the left-bank elevated plainland (the greater part of Sumy, Poltava, and Kharkiv regions).

The total area of the Forest-Steppe zone is 20 699 thousand hectares or 33.4% of the Ukrainian territory. The farmland occupies 35% of the state land fund, 65% of which is under plowland. The soil cover of the zone is very complex and represented by over 160 mapping units.

26.1. Factors of Soil Formation

The climate in the Forest-Steppe zone is moderately continental, warm with sufficient humidity in the west and unstable – in the east, with non-deficit balance of moisture, and periodic infiltrative type of water regime. The western province is the most humid in the zone. The hydrothermic coefficient exceeds 2. The annual precipitation equals 550-700 millimeters and even more; its main part – 400-500 mm – falls during the warm season of the year. Rainstorms over 100 mm are possible here. The available moisture stored in the 0-100 cm layer of soil in spring may reach 160-180 mm reducing to 50-80 mm in the crop vegetation period. The crops are well provided with heat. The sum of active temperatures over 10°C amounts to 2400-2600°C; the frostless period lasts 165-190 days.

Further to the east, the provinces of the Forest-Steppe zone become considerably less humid. The hydrothermal coefficient gradually reduces from 1.2 to 1.0. The annual precipitation is within 580-500 mm: 350 to 400 mm falls in the warm season of the year. The crops are slightly less well supplied with available moisture, the stores of which in spring amount to 130-170 mm in the soils upper layer of 1 m reducing to 50-80 mm at the harvest time. The sum of active temperatures amounts to 2600-2800°C and even up to 3000°C near the boundary with the Steppe zone in the east. The water balance remains positive in the western province, whereas on the other territory it keeps close to neutral. Precipitation roughly equals evaporation. Climate is an important factor of crop production. Forest-Steppe farms specialize in the production of winter wheat, corn, sugar beet, vegetables, fruits, berries, and forage crops.

The relief in the Forest-Steppe zone is quite diverse and specific in its various parts. Generally, the zone is an uplifted plainland with well-developed erosional landforms. The Right-Bank Forest-Steppe has wavy water-erosional forms of relief. Such a relief is typical for the old plateau territories with well developed streams patterns. The valleys are deep, wide, and long but not too branching. Stream patterns are dense. The watersheds between the valleys are elevated, wide, with well-expressed plateaus, the area of which is approximately equal to that of the slopes. The slopes are well expressed morphologically, mostly level, not too steep, but with increasing steepness near the valleys.

Watershed plateaus have a well-developed system of micro landforms like depressions and hollows.

The Right-Bank part of the Forest-Steppe zone is rich in low-productive lands eroded by gullies cutting their way into a landscape. 34.9% of the area in this region is an eroded land. Plowland is 31.4 percent eroded, and grasslands – 65.4%.

The Left-Bank Forest-Steppe is characterized by a mainly flat plainland topography. The landforms here are mostly erosional but stream patterns are less dense. The valleys and ravines are not deep, but wide, and often with wide and flat bottoms. The slopes of valleys are short, not steep and often with flat cross sections.

Surface flow is less rapid and energetic and water erosion is, therefore, weaker. But there are numerous poorly drained (isolated) depressions.

In Donetsk region, near the town of Siverodonetsk, on the high right bank of the Siversky Donets river, there is an area of dense network of gullies and ravines, veritable Ukrainian badland. In the Western Forest-Steppe zone, the land surface is somewhat wavy and occasionally with ridges, typical for characteristic of the elevated areas of ancient plateaus and the bands near the river banks. The stream pattern here is very dense with numerous narrow watersheds. The area of sloping land is greater than that of the plateaus. All forms of water erosion are evident here: sheet erosion, rill erosion, gully erosion, stream bank erosion, and sediment deposition.

Water-accumulating types of topography characteristic of the river terraces are covered with loess (or, rather, loess-like loams) and recent alluvial deposits. Such areas are flat plainlands where erosion is not evident. Owing to poor drainage, and accumulative land forms, ground water table is high and the soils are excessively wet.

The character of vegetation in the Forest-Steppe zone played (and still plays) a leading role in soil formation. A complicated evolution of vegetation is thought to determine the character of the soil cover in this zone, particarly the geography and topography of the most widely spread here podzolized and regraded chernozems (secondarily carbonatic or saturated with bases).

Forest and meadow-steppe vegetation areas alternate with each other. Forest vegetation is found mainly in elevated and well-drained areas and is represented by oak forests, as well as mixed oak-hornbeam and oak-maple-lime forests in the Right-Bank and the Left-Bank provinces, respectively.

In the Right-Bank part of the zone, the highest canopy in the forest is formed by oak, ash-tree, and sycamore and the lower canopy – by maple and hornbeam. The lowest canopy is formed by hazel-nut-tree, buckthorn, snowball tree, hawthorn, elder, etc. In the Left-Bank part of the zone, the highest canopy is formed by oak mixed with maple and lime. The bushes are represented by hazel-nut-tree, tartaric maple, hawthorn, and sweet briar. In the Western province the highest trees are represented by oaks, hornbeams, maples, sycamores, and occasionally by birch. The shrubs in the forests are represented by nut-trees, hawthorns, bird-cherries, etc.

The grasses in the forests are well developed and diverse: stellaria, strawberry, plantain, hoof-grass, lily of the valley, sedge, etc. Nettle, greater celandine, and chelidonium are widely spread weeds.

In the past, large areas of meadow steppes were characteristic of the Forest-Steppe zone. Nowadays the plant associations of such steppes can be found only in the Mykhailivska virgin land (Sumy region, Lebedin district), as wall such areas were put to plow.

Fescue grass, feather grass, bluegrass, and needle grass can be seen in the Mykhailivska virgin land. Cereal and leguminous grasses formed meadow-steppe associations of herbs. The amount of air-dry plant residues, both surface and root, annually left to decay, mineralize, and humify in the virgin meadow steppe was about 12-15 mt/ha.

Parent materials in the Forest-Steppe zone are characterized by "lithologic uniformity". With the exception of floodplain alluvial and terrace sandy soils, very eroded soils in the eluvium of solid rocks, and some soils in the lower parts of slopes, the rest of the soils were formed in loess and loess-like loams. The North-western part of the zone is characterized by the dominance of light loam loess with 50-60% of coarse silt fraction. Some sandy loams are also found there, as well as in the Central part of the Right-Bank province. Such texture instigates the development of gully erosion.

On the Left-bank terraces the loess is mainly medium loam, characterized by calcium and with neutral reaction which favor the formation of soils with good physical properties, water retention and penetrability, favorable physic-mechanical properties and soil structure. Loess contains 6-13% of calcium and magnesium carbonates. But the loess on the terraces, especially on the single-loess-layer terrace is saline or salt-affected mainly by sodium bicarbonate but also by such toxic salt as sodium carbonate the toxicity of which is ten times greater than that of sodium chloride. Sodicity-affected soils also form in such loess, the main cause for this process being ground water composition.

The Volyno-Podilska upland and the Left-Bank elevated plainland are characterized by the ancient water-erosional forms of topography. The localities are well-drained and with considerable erosion hazard. On noneroded plateaus, the thickness of loess deposits reaches 25-30 m. The deposits are divided into 5-6 layers by 4-5 horizons of buried soils.

In the southern Forest-Steppe a layer of "chocolate" loess appears the thickness of which grows to the south. The loess is of heavy texture (heavy loam or light clay).

26.2. Soil Genesis in the Forest-Steppe Zone

Forest-Steppe soils, with the exception of leached and typical chernozems, have a profile differentiated by eluvio-illuvial type. The leading processes of soil formation here are sod-forming (humuso-accumulative) and podzolic. The processes of lessivage, leaching, and regradation also take place. Grey forest soils, one of the zonal types, there are numerous hypotheses of the origin, of put forth by outstanding Russian pedologists, among them V.V.Dokuchaev, S.I. Korzhinsky, V.I.Tanfilyev, V.R.Williams, and I.V.Tyurin.

According to V.V.Dokuchaev, grey forest soils are a separate soil type. Light grey soils formed by the dominant effects of forest vegetation, and dark-grey by the more intensive influence of grasses. According to S.I. Korzhinsky, grey forest soils are of a secondary origin and were formed from chernozems by the attack of forest vegetation. This “attack” took place from the north. Forest vegetation ousted steppe one, causing the degradation of chernozems.

V.I.Tanfilyev, V.R.Williams, I.V.Tyurin and some other scientists thought differently. According to their inferences, the steppes attacked the forests from the south. Climate change substituted coniferous trees for broad-leaved ones. Podzolic process was slowed down. Sod formation became relatively more intensive. Grey forest soils developed into chernozemic ones. In places, were forests were preserved, so were grey forest soils.

V.V.Dokuchaev defended his idea of the genetic independence of grey forest soils. The recent research evidence and data on the biologic cycle of substances and the development of podzolic and sod-forming processes of soil formation prove the rightness of Dokuchaev’s view. The research shows that the geographic zone of grey forests soils is stable in space and in time. The soil formation under the canopy of broad-leaved, mainly oak forests with the participation of meadow-steppe grasses leads to the formation of grey forest soils, which should be considered an independent soil type.

The podzolic process under broad-leaved (deciduous) forests is less intensive than under coniferous ones. Well developed associations of grasses create favorable conditions for the development of the soddy process, which includes humus accumulation and its retention in situ to a considerable extent. Dead residues are richer in nitrogen, calcium, and other ash elements.

From the north to the south of the Forest-Steppe zone the forests become less thick-canopied and grasses in them grow better. Humus accumulation and the soil structure also improve.

Humuso-accumulative (sod-forming) process of soil formation is not self-sustained in the Forest-Steppe zone. It is combined with podzolization and lessivage during the formation of grey forest soils.

The podzolic process of soil formation in the zone is weakened, as forest litter in deciduous forests is rich in nitrogen and ash elements, especially calcium, and parent materials (loess) contain carbonates.

Lessivage is a mechanical transfer of colloidal particles in soil environment with low OM content, but quite biologically active and slightly to moderately acid. Organic acids are less aggressive in decomposing forest liter than in coniferous forests. So the mineral part of the soil is not subject to deep decomposition. Clay and colloidal particles are mechanically, without destruction, removed from the eluvial to illuvial portion of the soil profile. SiO_2 / R_2O_3 ratio in clay fraction of lessivated soils practically does not change with depth.

Regradation is another elementary, or first-order, soil formation process in the Forest-Steppe zone. Climate change, felling of forests, and turning the former forest land into a plowland, the hydrothermal soil regime also changes. The upward movement of water through the capillaries prevails over the downward movement. More water is evaporated than comes with precipitation. Deeply penetrating roots of the grassland vegetation absorb calcium from the lower soil horizons which remains in the soil after root residues mineralization. Periodic drying of the upper portion of soil profile proceeds according to the reaction:

$$Ca^{2+} + CO_2 + H_2O \rightarrow CaCO_3 + 2H^+$$

Soil adsorbing complex becomes more saturated with Ca^{2+} and the bulk of the soil – with $CaCO_3$. Previously leached soil horizons become carbonatic.

The origin of chernozems. V.V.Dokuchaev in his classical monograph *Russian Chernozem* (1883), after a critical review of all the existent hypotheses, came to the conclusion that chernozem was formed under the influence of grassland vegetation in any parent material and that neither forest nor swamp can form chernozems.

According to the recent concept, the chernozemic process of soil formation in the Forest-Steppe zone takes place under the canopy of perennial grassland vegetation in a moderately humid climate (hydrothermal coefficient equal to unity) mainly in loose carbonatic parent materials (loess and loess-like loams). The essence of the process is in the enrichment of parent material or the upper horizons of the solum by specific humus substances of acidic nature ($-COOH$ groups), consisting of humic, fulvic, and hymatomelanic acids, as well as humin (insoluble residue) with the dominance of humic acids. Humus synthesis and accumulation are the main constituents of the chernozemic process of soil formation.

Humus formation (synthesis) is the microbial decomposition of plant residues in situ and their subsequent humification without any transfer in soil profile. It is characterized by the formation of humus films with dark coloring on the surface and within the clods and grains of soil structure and by the generally darker color of the upper soil horizons contain one the greatest amount of lime and dead roots.

Humus accumulation occurs in the surface horizon as a result of decomposition of organic residues, their humification, transfer, and gradual enrichment of soil horizons with humus. The process leads to the formation of well humusified and structured soil horizons in the upper part of soil profile.

Moderate and periodic wetting of the soil is characterized by the alteration of downward and upward movement of soil water which favors the uniform saturation of soil horizons with humus. Calcium and magnesium carbonates may be leached from the upper to lower horizons or even parent material. Saturation of a soil absorbing complex with exchangeable Ca^{2+} and coagulation (flocculation) of soil colloids are the causes for the formation of waterproof cloddy and grainy aggregates. Humus accumulation is accompanied by the accumulation of N, S, P, K, Mg, Ca, and other macro-and microelements.

The chernozemic soil profile demonstrates a gradual transition from a highly humified, saturated with bases, and structured surface horizon (H) to parent material. There is no distinct redistribution of colloidal matter in the profile, especially in that of typical and leached chernozems. Under certain conditions, not without the effects of forest vegetation, humus accumulation is slightly impeded by podzolic processes. The result of such an interaction is the formation of podzolized chernozems.

Podzolized chernozems are thought to have gone through two phases of their development – lasting long under forest vegetation (podzolic) and a longer one – under steppe grasses (chernozemic). Their profile, therefore, has some, not too distinct, signs of podzolization – the presence of amorphous silica powder in the surface horizon (H_e), and compactness, brownish tint, and nutty structure in transitive horizons (H_{pi}, P_{hi}). Mole tunnels *crotovinas* (кротовини) can be found in any Forest-Steppe chernozem to the depth of 2 meters and more. The podzolized chernozem profile is more distinctly differentiated in the varieties with a heavier texture.

The complex interrelations of humus accumulation and leaching leads to the formation of leached chernozems.

Leaching, in the context of Ukrainian soil science, is the removal of $CaCO_3$ $(MgCO_3)$ from carbonatic (effervescing with 10% HCl) soil layers. The soil may be leached from carbonates during the periods of heavy rainfalls. Not too heavy texture also favors leaching: $CaCO_3 + H_2O + CO_2 \rightarrow Ca(HCO_3)_2$. Leached chernozems were formed under thinned out deciduous forests or on the peripheral parts of forest areas under meadow grasses. Leaching of carbonates and partial loss of exchangeable Ca^{2+} reduced SOM content in the upper horizons and increase the overall depth of the horizons with SOM accumulation.

Podzolized chernozems, like the grey forest soils, may undergo regradation, and formerly leached and illuviated or even eluviated horizons, may become carbonatic. Light grey and grey forest soils are found in the northern part of the Forest-Steppe zone. Their area is about 1, 635 thousand hectares (12.4%). Podzolized soils (dark grey soils and podzolized chernozems) occupy an area of 3, 271 thousand hectares (24.7%), and the largest area (7, 214 thousand hectares, 54.6% of plowland) is occupied by typical chernozems.

26.3. Grey Forest Soils and Their Classification

N.B. Vernander, N.M. Breus, M.I. Polupan and other scientists, taking into account the intensity and results of the podzolic process in the Forest-Steppe zone, proposed to identify two well-expressed soil types: grey forest soils and podzolized ones. The classification of grey forest soils, including the subtypes of grey proper and light grey forest soils is given in Table 70.

Grey and light grey brownish and wet (or just humid) soils form in the Western province of the zone. In the right, and to a lesser extent in the Left-Bank parts of the zone. Grey and light-grey soils form Grey forest mountain soils occur in the mountain region of the Crimea.

Table 70. Classification of Grey Forest Soils (Research Institute of Soil Science and Agrochemistry, 1988)

Type	Subtype	Genus	Species
Grey forest soils	Light grey brownish, light grey wet, light grey, grey brownish, grey wet, grey, and grey mountain	Modal, eroded, secondarily meadowish, regraded, mocharic, plantaged (deeply plowed)	Modal, by the extent of erodedness, regradedness, depth of meadowish effects and underlying
Grey forest soils, gleyed from the surface	Light grey and grey – both gleyed from the surface	Modal	By the extent of gleying
Grey forest gleyed soils	Light grey and grey, both gleyed	Modal, saline	Modal, by the depth of salinity and gleying, and extent of their development

The soils are grouped into light grey and grey by morphologic features. The former have a humuso-eluvial horizon up to 20 cm deep, and a well- expressed eluvial horizon (E_h), whereas the latter have a deeper HE horizon (20-30 cm) and their eluvial horizon is not clearly expressed, or it may be absent and I horizon follows the HE.

The most widespread genera of grey forest soils are:

1) **modal (ordinary) soils.** As the name implies, they have no outstanding features;

2) **washed-on soils.** They are covered with erosion sediment from the surrounding slopes, their HE horizon is deeper and often layered;

3) **regraded** (secondarily saturated) soils. Effervescence with 10% HCl occurs in illuvial (I_k) or, less often, in EI and HE horizons ($EI/_k$, $HE/_k$);

4) **meadow and secondarily meadowish soils**. They are found in the areas of waterlogging among the grey forest soils. The meadowish ones have ground water

table 3-5 m deep, and meadow – up to 3 m deep. Parent materials may be gleyed or metamorphosed (marled);

5) **saline soils.** They form in the areas with elevated water table of saline water among the grey forest soils.

Grey forest soils are divided into species by the extent of erosion they suffer from:

1) **slightly eroded**: HE horizon is lost by 1/3 of its initial thickness, the loss of SOM is up to 20%, rills may occur on the surface of a plowed field;

2) **moderately eroded:** over 1/3 of the HE horizon is lost; the loss of SOM is within 20-40% of its initial content; the upper part of the I_h horizon is involved into a plow layer, which has a brownish tint;

3) **severally eroded:** the entire HE horizon is washed off, as well as EI and partly I horizons; SOM losses are within 40-60% of its initial content; the plow layer consists mainly of illuvial horizon.

By the extent of regradation, the following species are identified:

1)**slightly regraded,** when the slightly illuviated parent material is carbonatic (P_{ik}) or the lower illuvial horizon (PI_k);

2)**moderately regraded,** when a middle part of the illuvial horizon contains carbonates (I_k); and

3)**very regraded,** when the upper portion of the illuvial horizon contains carbonates (I_{hk}).

By the character and extent of reduction processes (gleying), the following species are identified;

1)**gleyed from the surface:** they are characterized by the surface gleying of the entire profile and the upper part of the noncarbonate parent material. Such species are not identified among the grey forest soils;

2)**gleyed from below:** the wetting of the soil is mixed, but gleing is caused mainly by high (2-4 m) watertable. Such gleying is identified only in grey forest gleyed soils, which are divided into gleyed and gleyish;

3)**gleyish:** they have the signs of gleying in illuvial horizons transitive to the parent material and in the parent material (IP_{gl}, Pi_{gl}, P_{kgl}); signs of gleying in the form of dull grey or rusty-ochric spots and extended patches; sometimes beans of Fe and Mn concretions are found too;

4)**gleyed:** they manifest signs of gleying in the illuvial, transitive illuvial horizons of the solum and in the parent material (I_{gl}, I_{pGl}, PI_{gl}, P_{kgl}). The signs of gleying are more distinct than in the previous species, and dull grey and greenish colors are brighter. Sometimes beans of Fe and Mn concretions are found not very often.

The species identified by the depth and extent of salinity are:

1)**deeply saline,** salted mainly with sodium biocarbonate from the depth of 40-50 cm, they occur only among gleyed grey forest soils,

2)**saline all over the profile** (solonchaky), which also is salted mainly with sodium biocarbonate. The soil reaction is slightly alkaline. Crystalls of salts and even crusts of salt may appear on the surface when the soil dries up. Such species are identified only within the genus of gleyed grey forest soils.

A lithologic series refers to the parent material of soil, which is dominantly loess (loess-like loam). But nonloess parent materials may also be found, like alluvial, glaciofluvial, red clay and other deposits underlain by sands, pebbles, limestones, schists, etc. Among the **variants,** there are virgin, cultured, irrigated, and drained ones.

26.4. Grey Forest Soils Morphological Features

26.4.1. Grey Forest Soils

Genetically, the profile of grey forest soils is distinctly differentiated into eluvial and illuvial parts. The surface horizon is “responsible’ for humus accumulation. It is usually followed by an eluvial-illuvial horizon, which, in its turn, is followed by an illuviated parent material. Light grey and grey forest soils are differentiated as subtypes.

Light Grey Forest Soils are widespread mainly in the Central Forest-Steppe province. They lie on well drained territories and are surrounded by less podzolized soils. On the Left-Bank province these soils occur as relatively small areas within the mapping units of grey forest soils on high, well drained terraces or as “inclusions” on the “loess islands” of the Forest zone (Polissya). Morphologically, these soils look very much like soddy-podzolic ones but they were formed in loess and loess-like loams containing $CaCO_3$. Virgin variants (under the forest) have a considerable humus content (4-5.5%), whereas plowland variants have been “cultured” to such an extent, that their humus content amounts to 1-2.3%. Fulvic acids dominate in soil-humus. Below is a brief description of a light grey forest soil:

H_0: forest litter; 1-2 cm thick; formed of well decomposed leaves and branches;

HE: humuso-eluvial horizon; in virgin variants – 8-15 cm thick; in arable variants forming a plow layer 26-28 cm thick; light grey, poor in clay, with amorphous SiO_2

"dusting" becoming more distinct on drying; powdery and cloddy aggregates dominating over plates; slightly compacted; transition clear.

$E_{(h)}$: eluvial horizon; very slightly and nonuniformly humusified; under forest – 8-15 cm thick; in the field its thickness is only 1-2 cm, as it was involved into a plow layer; whitish brown, rich in amorphous silica; platy or leafy; transition clear;

I: illuvial horizon; 35-60 cm thick; reddish brown or brown; rich in R_2O_3; prismatic; amorphous SiO_2 on the faces of aggregates, as well as "colloidal lustre"; compacted; transition gradual;

IP: transitive to the parent material well illuviated horizon; 15-35 cm thick; red-brown; prismatic-cloddy; less rich in R_2O_3 and less compacted than the previous horizon; transition gradual;

P_i: slightly illuviated parent material (actually, just leached from carbonates); 20-30 cm thick; brownish dull yellow; rare layers of R_2O_3; cloddy; slightly compacted; transition distinct, along the line of effervescence from 10% HCl;

P_k: parent material; dull yellow (straw-colored); with veins or molds of $CaCO_3$.

Grey forest soils are widespread in the Central Forest-Steppe zone. They occupy large areas surrounded by podzolized soils (dark grey soils and podzolized chernozems). On the left-bank they occur on high loess terraces of the Dnipro's tributaries. In Polissya zone they occupy "loess islands". Forest (virgin) soils contain 4-6% of SOM in the surface horizon (HE), whereas in the field their humus content is within 1.5-2.7%. Fulvic acids dominate in soil humus. One of the profiles had the following description:

H_0: forest litter, 1-2 cm thick, composed of decaying leaf and branch residues;

HE: humuso-eluvial horizon; 25-35 cm thick; darkly grey; grainy with clods, or, more often, cloddy with grains; sometimes with platy "cleavage", with abundant SiO_2 on the faces of soil aggregates; slightly compacted; transition clear;

EI_h: eluvio-illuvial, slightly and nonuniformly humusified horizon; 15-20 cm thick; dirty brown; spotty; whitish spots of amorphous SiO_2; streaks of humusified material; large-nutty; compacted; red-brown "varnish" on the faces of aggregates; transition clear;

I: illuvial horizon; 40-60 cm thick; red brown from the abundance of R_2O_3; prismatic; colloidal films on the faces of soil aggregates; compacted; transition gradual;

IP: transitive to parent material illuviated horizon; 40-60 cm thick; reddish brown; rich in R_2O_3; prismatic; colloidal films on the faces of soil aggregates; compacted; transition gradual;

P$_i$: slightly illuvial parent material; 20-30 cm thick; dull yellow with a brownish tint; cloddy; slightly compacted; transition distinct (along the line of effervescence with 10% HCl).

P$_k$: parent material; dull-yellow (straw-colored) loess; visible carbonates in the form of veins and pseudomicellium.

26.4.2. Brownish Grey Forest Soils

Brownish grey forest soils soils occur in small areas in Prycarpathia, Opillya, Rostochchya, Podillya, and between the Prut and the Dnister. They formed under deciduous forests consisting of beech and elm, beech and hornbeam, and oak and hornbeam; with well developed grass cover in loess-like parent materials. With good drainage and humid climate, grey forest soils acquire the features of brown forest soils: profound leaching, unsaturation with bases, considerable acidity, and over average content of available Al. The soil profile is differentiated by eluvium-illuvium type, the sign of colloidal transfer being expressed rather weakly, and with weak colloidal “luster” and the formation of clay in the illuvial portion of the profile.

Light grey brownish soils and grey brownish soils are two subtypes within the type. Light grey brownish soils profile is composed of the horizons: $H_{0\ 1\text{-}2}$ + $HE_{\ 10\text{-}18}$ + $E_{(h)\ 10\text{-}15}$ + $I_{\ 70\text{-}90}$ + $P_{i\ 20\text{-}40}$ + P.

Grey forest brownish soils spread among the light grey forest brownish soils. Their formation and profile composition is complex. They do not have any eluvial horizon [$E_{(h)}$] and their HE and I horizons are better developed and better humusified: $H_{0\ 1\text{-}2}$ + $HE_{\ 20\text{-}35}$ + $I_{h\ 50\text{-}70}$ + $P_{i\ 30\text{-}50}$ + P.

26.4.3. Grey Forest Wet Soils

Grey forest wet soils they occur in the Western Forest-Steppe zone, but unlike brownish soils, they were formed in parent materials of heavier texture.

High rainfall and low water penetrability of soils cause periodic wetness and seasonal gleying which shows itself in the form of rusty brown spots, Fe and Mn concretions (beans and pointations), the content of which in the layer immediately overlying illuvium may reach 12% of soil weight. Carbonates accumulate in the form of large concretions ("dolls"). The soils are characterized by deep leaching and high immovability of clay particles which fill in the cavities, earthworm tunnels, cracks, etc. The soil profile (solum) may reach the depth of 3-4 meters and more. Humus content is within a wide range of 1.5 to 5%. Fulvic acids dominate in soil humus. Soil reaction is acid as soil solution contains 10-15 milliequavalents of Al^{3+} per 100 g of oven dry soil. PBS is within 65-75%.

Wet light grey forest soils are widespread in Rostochchya, Opillya, and between the Prut and the Dnister in the Western Forest-Steppe province. They form on large and medium ridges and high loess terraces where they lie on watersheds and their slopes. Ground waters are over 6 meters deep and do not effect soil formation. The soils were formed under oak-beech and beech-hornbeam forests, with well developed grass cover, in loess and loess like heavy loams underlain by clays, slates, and limestones, practically impenetrable to water. The profile of the soil is composed of the horizons: $H_{0\ 1\text{-}2} + HE_{gl\ 28\text{-}30} + E_{gl\ 13\text{-}17} + I_{gl\ 60\text{-}90} + IP_{gl\ 25\text{-}35} + P_{igl\ 20\text{-}40} + P_{kgl}$.

Wet grey forest soils occur in the same areas as light grey ones. Humus accumulation in their profiles is better, and their differentiation – less distinct. Their profile may be given as: $H_{0\ 1\text{-}2} + HE_{gl\ 30\text{-}45} + I_{gl\ 70\text{-}80} + IP_{gl\ 30\text{-}40} + P_{igl\ 30\text{-}50} + P_{kgl}$. As can be seen, the gley (gl) horizon is absent.

26.4.4. Grey Forest Gleyed Soils

Grey forest gleyed soils ocur in the Forest-Steppe and Polissya zones where they form on poorly drained watersheds and loess terraces, in saucer-shaped depressions which suffer from ponding and waterlogging due to a high ground water table. The periodic washing of the entire solum with water and partial anaerobiosis resulting from it cause the development of gleying. Soil formation products are partly leached beyond soil horizons. The soils, therefore, are more eluviated, more acid, and less saturated with bases in comparison with automorphic soil units. The layer of SOM accumulation is also thicker. Grey forest gleyed soils are formed in the negative elements of topography, whereas grey forest wet soils form on the watersheds.

There are two subtypes of given soils: gleyed from the surface and gleyed from the depth. The first subtype has the following profile: $H_{d\ 2\text{-}5} + HE_{gldl\ 20\text{-}25} + E_{igl\ 8\text{-}15} + IE_{gl\ 15\text{-}20} + I_{egl\ 30\text{-}50} + IP_{egl15\text{-}25} + P_{iegl\ 20\text{-}40} + P_{gl}$.

The second subtype's profile is composed in the following way: $H_{d\ 2\text{-}5} + HE_{gldl\ 30\text{-}40} + IE_{gl\ 40\text{-}50} + I_{egl\ 30\text{-}50} + IP_{egl\ 20\text{-}30} + P_{igl\ 30\text{-}40} + P_{gl}$. The soils are better supplied with SOM as a result of deluvial deposition from the surrounding areas.

26.4.5. Composition and Properties of Grey Forest Soils

Grey forest soils form mainly in loess and loess-like parent materials (97%). Their texture varies from loamy sand to heavy loam. Light (42.9%) and medium (34.3%) loams are the most widespread ones; loamy-sands occupy only 16.8 and heavy loams – only 6% of their total area. Down the profile, the soil texture becomes heavier, especially in the illuvial horizon.

The chemical composition of different horizons is different. The maximum SiO_2 content (81-85%) is characteristic of HE horizons, whereas oxides of iron and aluminum (3-4% and 9-12%, respectively) accumulate in the illuvial horizon. SiO_2 / R_2O_3 ratio is the lowest (9-12) in the illuvial horizon. This is the evidence of a podzolic process taking part in soil formation.

SOM (humus) content in light grey forest soils is very low (1.19%); it is higher in grey forest soils (2.03%), but in both soils it sharply decreases with depth (Table 17.2). The amount of SOM in the 1-meter layer of soil (mt/ha) in light grey soil may be very low (87) to low (125), whereas in grey soils it may be very low (93) to moderate (241). The composition of SOM is only slightly differentiated in the profile. But it may be humate-fulvatic in the HE and fulvatic in the I horizons.

Fulvic acids are bound into fractions with labile R_2O_3, but some of them are free and aggressive ruining the mineral part of the soil and migrating to the illuvial horizons, where their content is the highest (58-75%). Physico-chemical properties of grey forest soils reflect their genesis. By exchangeable acidity (pH_{KCl} 4.4-5.5), they are moderately acid. Soil solution is slightly acid (pH_{H2O} 5.1-5.3). Their hydrolytic acidity is about 3-4 meq. per 100 g of oven-dry soil. Down the profile the soil acidity decreases (Table 71).

Calcium dominates among exchangeable cations. PBS shows the soils to be moderately and well saturated with bases (64-88%). Parent materials contain $CaCO_3$ (2.7-7.8%). Judging by physico-chemical characteristics, the intensity of podzolic process in grey forest soils is quite high, but it is restricted by the grass cover in the forests and presence of $CaCO_3$ in parent materials.

Low SOM content makes the soils poor in total nitrogen, which also depends on soil texture. Thus, in the HE horizon of light loams the total nitrogen content is within 0.09-0.11%, of medium loams – 0.12-0.16, and of heavy loams it is 0.14-0.19%.

Table 71. Content and Group Composition of Humus in Grey Forest Soils (D.I. Kovalishin and N.M. Breus, 1979)

Soil horizon and depth, cm	Total SOM content, %	C org. content, %	C h.a.	C f.a.	C residual	C h.a./ C f.a.
			% of C org.			
Light Grey Forest Soil in Light Loam Loess						
HE_{2-12}	1.19	0.69	20.6	22.6	56.8	0.91
$E_{h\ 22-33}$	0.94	0.55	20.0	27.2	52.8	0.73
$I_{(h)\ 41-51}$	0.71	0.41	10.7	42.9	46.4	0.25
I_{85-95}	0.32	0.19	10.5	57.9	39.6	0.18
$P_{k\ 140-150}$	0.33	0.19	10.0	58.9	21.1	0.34
Grey Forest Medium Loam Soil in Loess						
HE_{10-20}	2.03	1.18	25.9	40.9	33.2	0.63
$EI_{(h)\ 30-40}$	0.69	0.40	30.8	52.0	17.2	0.59
I_{75-95}	0.43	0.25	32.8	74.8	not determined	0.43

Due to the soil acidity and not too favorable water and aeration regimes, the soils nitrification energy is low, and they are poorly supplied with mineral forms of nitrogen. Agricultural crops suffer the deficiency of nitrogen and respond well to its application.

The total phosphorus content is also low. It depends on the soil texture and does not exceed 0.10-0.13%, so phosphorus deficiencies are also common in such soils.

The soils are better supplied with total potassium, the content of which also depends on the soil texture. Thus, light grey soils contain 1.47-1.62% of total potassium in the HE horizon, and grey soils of a slightly heavier texture – 1.62-1.74%. This is approximately twice as much as in soddy-podzolic soils.

Table 72. Physico-Chemical Characteristics of Grey Forest Soils (D.I. Kovalishin and N.M.Breus, 1979)

Characteristics	Light Grey Light Loam Soil in Loess					Grey Medium Loam Soil in Loess				
	Soil Horizons									
	HE	E_h	$I_{(h)}$	I	P_k	HE	$EI_{(h)}$	I	P_i	P_k
Layer, cm	2-12	22-23	41-51	85-95	140-150	10-20	30-40	75-95	125-135	170-180
pH_{H2O}	5.1	4.2	4.8	4.8	7.4	5.3	5.4	5.2	5.6	7.6
pH $_{KCl}$	4.4	Not determined	4.1	4.1	Not determined	4.5	4.8	4.3	4.5	Not determined
Adsorbed cations, meq per 100 g of dry soil:										
Ca^{2+}	12.1	4.61	4.26	9.33	Not determined	15.0	14.0	10.8	19.7	Not determined
Mg^{2+}	3.79	2.10	3.23	4.23	>> – >>	4.0	1.55	6.89	4.91	>> – >>
Na^+	0.26	0.17	0.31	0.43	>> – >>	0.20	0.63	0.80	0.34	>> – >>
K^+	0.36	0.21	0.31	0.38	>> – >>	0.21	0.27	0.57	0.48	>> – >>
S	16.5	7.19	11.1	14.4		20.0	16.4	19.1	16.4	
Hydrolytic Acidity, meq/100g	4.1	4.0	4.2	3.2	>> – >>	3.52	2.23	3.34	2.42	>> – >>
CEC, meq per 100 g of dry soil	20.6	11.2	15.3	17.6	>> – >>	235	18.6	22.4	18.9	>> – >>
PBS, %	80	64	72	82	>> – >>	85	88	85	87	>> – >>
$CaCO_3$, %	Not determined				7.73	Not determined				2.68

The soils are moderately supplied with available P_2O_5 and exchangeable K_2O. The crops usually respond well to the application of both phosphorus and potassium fertilizers. The particle density of grey forest soils increases with depth (from 2.61 in HE to 2.72 g/cm^3 in the parent material). By the bulk density of the soil in the HE horizons (1.35-1.36), the soils are moderately compacted, whereas all soil horizons below HE are compacted (1.46-1.54 g/cm^3). As a result the total porosity is unsatisfactory (48-49% of volume in the HE and 43% in the illuvial horizon). Capillary pores dominate over non-capillary (large) ones. Poor aeration favors the development of reduction processes.

Minimal values of PWP (permanent wilting point) are characteristic of humuso-eluvial horizons (5.5-5.8%). In illuvial horizons the PWP increases to 10.5-13.6%. Relatively high field capacity (FC, 19-26%) allows the soils to have a sufficient potential amount of available water which in soils of different genesis and texture remains within 150-200 mm in the 1 m layer.

The soils may have low infiltrability, and be prone to puddling and crust formation. That is the reason why the water regime and the soil aeration are not always good in grey forest soils.

By potential productivity (bonitet) evaluation, according to A.I.Siry, the light grey forest soils may be given 30-40 points on the 100-point scale, whereas grey forest soils may be given 33-49 points.

26.4.6. Agricultural Use and Practices of Grey Forest Soil Improvement

Light grey and grey forest soils are generally good for crop production. With proper (contour) organization of fields, optimal tillage systems, good fertilizing, chemical amendments, plant protection, and crop rotations the soils are moderately favorable for the production of winter wheat, rye, spring small grain crops (barley, buckwheat, millet), corn for grain and silage, sugar beet, potatoes, clover, alfalfa, etc. They are also favorable for the production of fruits and berries.

Low humus content, high acidity, not sufficient content of available nutrients, physical properties (with an opportunites) for improvement, erosion hazard, and unstable water regime – all these factors deternine the set of practices for soil improvement.

The soils need subsoiling. But this practice of tillage can be a success only if sufficient rates of organic manures are applied. Farm manure is applied under row crops at rates of 30-40 mt/ha. To keep a non-deficit SOM balance, 16-18 mt of FM should be applied per hectare of a crop rotation per year. The use of composts is a way out of a complicated situation with manure shortage.

Rates of liming are computed as complete rates by hydrolytic acidity. Mineral fertilizers should be applied when soil tests show the macro- and microelements of plant nutrition to be below optimum (deficient). The rates should be as required by the fertilizing systems taking into account concrete yield goals for concrete crops. Beside superphosphate, phosphorite meal may also be applied, the best results of which may be achieved with no deficiency of nitrogen.

Minimum or restricted tillage may be a success provided that the primary tillage operations for different crops are performed to different depths. A non-plow tillage (with no furrow inversion) performed to varying depths may prove economically feasible and effective in soil and water conservation. Some researches, though, support the idea of combined plowing – non-plowing tillage. Additional surface cultivation is needed to destroy the crust.

Soil protecting crop rotations without row crops with non-plow primary tillage to differing depths should be employed on sloping lands within 3-5° of steepness. Tillage should be directed along the contours.

26.5. Podzolized Soils

This type of soils unites dark grey podzolized soils and podzolized chernozems. They occur mainly in the northern, central and western parts of the Forest-Steppe zone. In the Left-Bank province they occupy some areas on the right banks of the rivers.

According to N.M. Breus (1988), the area of podzolized soils used in agriculture is about 5,528 thousand hectares (13.8% of the total farmland area). The diversity of podzolized soils increases at the expense of regradation (27.8%), erodedness (36.5%), wetness (10.5%), drainage (5.8%), etc.

Dark grey podzolized soils almost never form large unbroken areas. As a rule, they are found among podzolized chernozems and grey forest soils.

Podzolized chernozems occur mainly in the western regions, the central part of Kyiv oblast, eastern periphery of the Volyno-Podilsky plateau, in Vinnytsya oblast and in the northern parts of Chernigiv and Poltava oblasts. They do not form an unbroken band but are invading the areas of typical and leached chernozems, as well as dark grey soils on watersheds and gentle slopes. The natural productivity of dark grey soils and podzolized chernozems is approximately of the same level and they are united into a single agroproduction group. Podzolized soils occupy 24.7% of a plowland in the Forest-Steppe zone.

26.5.1. Podzolized Soils Classification

Dark grey podzolized soils are characterized by a considerable accumulation of SOM, profile differentiation into eluvial and illuvial parts, unsaturation of soil adsorbing complex with exchangeable calcium, and a slightly acid soil reaction.

They are divided into three types and five subtypes: dark grey podzolized brownish, wet, modal, gleyed from the surface and gleyed proper (Table 73).

Table 73. Classification of Podzolized Soils (Research Institute of Soil Science and Agrochemistry, 1988)

Type	Sybtype	Genus	Species
Podzolized	Dark grey brownish, dark grey wet, dark grey proper, podzolized chernozems, brownish, podzolized chernozems wet, podzolized chernozems proper	Modal. Eroded, secondary meadowed, regrated, mocharic, deeply plowed	Modal, by the extent of erosion, regarding, depth of meadowishness and underlying
Podzolized gleyed from the surface	Dark grey gleyed from the surface, podzolized chernozems gleyed from the surface	Modal	By the extent of gleying
Podzolized gleyed	Dark grey podzolized	Modal, saline	Modal, by the depth of profile, extent of gleying and salinity

The most widely occurring genuses of dark grey podzolized soils are:

1) **modal (ordinary):** with typical subtype characteristics, the word *modal* is not used in the name;

2) **eroded:** may be characterized by partial or complete loss of upper horizons as a result of erosion;

3) **washed-on:** formed at the expense of deluvial deposits on the lower parts of slopes and valley bottoms, have a deeper HE horizon which is platy;

4) **regraded:** differ from the modal genuses by the presence of carbonates in illuvial horizon or over it;

5) **meadow and secondarily meadow:** are found in the areas of waterlogging. Ground water table in meadowish soils are 3-5 m deep; in meadow ones – it is less than 3 m deep. Parent materials, depending on the extent of waterlogging, may be gleyed or "marly";

6) **saline:** form in places with high (2-4 m from the surface) ground water table with saline water.

By the severity of soil erosion, the following species of brownish dark grey podzolized soils are identified as follows:

1) slightly eroded: up to ½ of the HE horizon and up to 26% of soil humus lost. Plowland is dark grey with brown tint, as a part of the HI horizon it is involved in a plow layer such soils, form on gentle slopes;

2) moderately eroded: the HE horizon is lost completely, as well as 20-40% of SOM. Plowland is brown-colored;

3) severely eroded: the soil lost its SOM accumulation horizon (HE + HI) as well as upper part of the I horizon, and 40-60% of its initial content of SOM. Plowed field looks brown, with large clods and rills 0.5-1 m deep. Such soils usually occupy steep convex slopes.

Eroded species of wet and proper dark grey podzolized soils are identified by the analogous criteria, so we do not describe them here.

By the extent of regradation the species are identified in the following way:

1) **slightly regraded:** effervescing with 10% HCl are the slightly illuviated part of the parent material (P_{ik}) and the lower portion of illuvial horizon ($I/_{k}$);

2) **moderately regraded:** the entire illuvial horizon (I_k) contains carbonates;

3) **very regraded:** the lower portion of humuso-illuvial horizon contains carbonates ($HI/_{k}$).

The species identified by the character and extent of gleying are as follows:

1) **gleyed from the surface:** are characterized by gleying from the surface of the solum and the upper portion of leached from carbonates parent material. Signs of gleying are in the form of ochre spots, beans, and pointations; not identified in dark grey wet and gleyed soils;

2) **gleyed from the depth:** sources of wetness combine with each other or gleying is caused by the high ground water table (2-4 m deep). They are identified only in dark grey podzolized gleyed soils subdividing them into gleyish and gleyed:

1) **gleyish soils,** which have P_{gl} or P_{Gl} and P_{igl}; signs of gleying are expressed weaker (dull grey, rusty, or ochrespots, and rare beans of Fe-Mn concretions);

2) **gleyed soils,** which have gleying in parent material, transitive illuviated and illuvial horizons (P_{gl} + P_{igl} + IP_{gl} + I_{gl}). Signs of gleying are very expressive (dull-grey or greish-green color, abundant concretions, etc). The parent material may be so sticky and looking like marl that it is designated as P_{Gl}.

By the depth and extent of soil salinity, dark grey podzolized gleyed soils are divided into two species:

1) **solonchaky (saline) from depth** (which is 40-50 cm);

2) **solonchaky (saline) from the surface**. Sodium bicarbonate is the main salting salt, so the dry soil shows signs of effervescence with 10% HCl.

According to soils classification, there are lithologic series, variants, and varieties of soils, analogous to those of grey forest soils.

Podzolized chernozems have the following genuses:

1) **modal** (ordinary, typical) the soil features of which correspond to the essential subtype characteristics. The word *modal* is not used in the soil name;

2) **eroded,** which differ from the modal ones by partial or complete absence of the upper horizons as a result of water erosion;

3) **regraded,** which differ from the modal genus by the presence of carbonates in the solum;

4) **gleyed from the surface,** caused to be gleyed by the surface water;

5) **secondarily meadowish and meadow,** which are encountered in the zones of waterlogging. Meadowish species have ground water 3-5 m deep, whereas meadow ones have it less than 3 m deep and are more severely gleyed.

Podzolized chernozems are divided into species by the extent of soil erosion:

1) **slightly eroded**: up to 1/2 of the He horizon and up to 20% of SOM stocks lost;

2) **moderately eroded:** He and partly Hpi horizons are washed off; 20-40% of SOM are lost from the soil;

3) **severely eroded:** H_e and H_{pi} horizons and 40-60% of SOM are lost from the soil.

By the extent of regradation podzolized chernozems have the following species:

1) **slightly regraded:** soil profile contains the horizons $P_{(i)k}$ and $P_{h(i)k}$;

2) **moderately regraded:** soil horizon H_{pik} and the horizon below it contain carbonates;

3) **very regraded:** soil horizon He/k in its lower portion and the horizons below it contain carbonates.

Podzolized chernozems are divided into species according to the depths of the layer of SOM accumulation (H_e + H_{pi}): not deep (45-65 cm), moderately deep (65-85), deep (85-120 cm), and very deep (over 120 cm).

By the content of SOM there are the species of slightly humusified (less than 3%), lowly humusified (3-6%), and moderately (averagely) humusified (over 6%) soils. Any soil surveyor should be able to identify slightly and moderately podzolized soils in the field.

26.5.2. Podzolized Soils' Morphological Features

A schematic description of a dark grey podzolized soils profile may be like this:

Hd (H0): forest litter or sod, 2-4 cm thick;

HE: humuso-eluvial horizon; 30-35 cm thick, dark grey, rich in amorphous SiO_2, grainy with clods in a plow layer, platy in the layer of a plow toe; the transition to the underlying horizon – clear;

HI: humuso-illuvial horizon; 30-35 cm thick, dark grey with brownish tint, nutty, rich in R_2O_3, compacted, with rare earthworm tunnels and holes made by roots; transition – clear;

I: illuvial horizon; 30-50 cm thick, red brown or brown, prismatic, colloidal films of reddish brown color on the faces of aggregates, compacted, very rich in R_2O_3; transition – gradual;

IP: transitional horizon; well illuviated, 15-25 cm thick, reddish brown, cloddy with prisms, rich in colloidal matter, less compacted, transition – gradual;

P_i: slightly illuviated parent material; 20-30 cm thick; dull yellow with brown tint, layers of R_2O_3, cloddy, slightly compacted, transition – very distinct (along the line of effervescence with 10% HCl). Visible carbonates in the form of veins.

Brownish types of dark grey podzolized soils are characterized by a deep HE horizon and considerable SOM content (5-7.5% in the forest and 2.8-3% in the field). The HI horizon is extended, not distinct, rich in OM. The soil profile may be composed of the horizons: $H_{d\ 2\text{-}3\ (H0)}$ + $HE_{35\text{-}50}$ + $HI_{25\text{-}40}$ + $I_{50\text{-}90}$ + $P_{i\ 20\text{-}40}$ + P_k.

Wet types of dark grey podzolized soils are widespread in the Western Forest-Steppe zone. They have a deeper and better humusified profile. Podzolization and seasonal gleying features are not distinctly expressed. Virgin soils may contain 2.7-7% of SOM in the surface mineral horizon, whereas soils in the field – only 1.5-3.2%. The soil profile has the horizons: $H_{d\ 2\text{-}4}$ + $HE_{(gl)\ 30\text{-}45}$ + $HI_{(gl)\ 20\text{-}50}$ + $I_{(gl)\ 50\text{-}60}$ + $PI_{(gl)\ 25\text{-}35}$ + Pi(gl) 30-40 + Pk(gl).

Dark grey podzolized soils gleyed from the surface occur in the Forest-Steppe and Polissya zones on the negative elements of topography. Periodic puddling with low watertable (over 6 m from the surface) causes the development of gleying. The soil profile consists of the horizons: Hd 3-5 + Hegl 20-50 + HIgl 25-40 + Iegl 30-50 + PIegl 15-25 + Piegl 20-40 + Pgl.

Dark grey podzolized gleish soils cover poorly drained watersheds and terraces with high ground water table (within 2-4 m from the surface). The soil profile combines the horizons: Hd 3-5 + He 32-45 + HI 20-30 + IP 35-50 + Pigl 20-30 + Pkgl.

Dark grey podzolized gley soils are found on the same elements of topography, but have a deeper profile of SOM accumulation (He + Hi = 70-80 cm). Their profile is more gleyed: Hd 3-5 + He 35-50 + Hi 30-40 + Igl 35-50 + IPgl 30-50 + Pigl 25-35 + Pkgl.

Podzolized chernozems have a deep humusified profile and are encountered in the Right-Bank Forest-Steppe zone covering level poorly drained watersheds and their slopes among the dark grey podzolized soils, leached and typical chernozems.

Podzolized chernozem profile has the following morphological features:

HE (He): slightly to moderately eluviated horizon of SOM accumulation; 35-45 cm in depth; dark grey, displaying whitish dusting of amorphous SiO_2 on drying; plow layer cloddy with powdery grains; a layer below it – nutty with grains and clods; slightly compacted; transition gradual.

Hpi: upper transitive slightly illuviated horizon, 30-40 cm thick; dark grey; cloddy with nuts; compacted; with a weak amorphous silica dusting; brownish tint from the colloids of R_2O_3, rare earthworm channels and “crotovinas” (mole tunnels); transition gradual;

Phi: lower transitive slightly illuviated horizon; 35-45 cm thick; grey brown; nutty-prismatic; with R_2O_3 coatings (films) on the faces of soil aggregates; transition gradual;

Pi(h): slightly illuviated or just leached from carbonates horizon; 15-30 cm deep; with streaks of humus; dirty dull-grey with brownish tint; rare “crotovinas”; sharp transition along the wavy line of effervescence with 10% HCl;

P_k: parent material; straw-colored loess with veins of $CaCO_3$.

Regraded chernozems are more popular in the Right-Bank province in Cherkasy, Vinnytsya, Odesa, and Kirovograd regions. They are much less common in the Left-Bank province. Some areas of such soils are located between the Zbruch and the Seret in the Western Forest-Steppe zone. The soils are characterized by a higher content of SOM (4.5-8% under the forest and 3-5% in the fields). Humus type is fulvate-humatic (C_{ha}: C_{fa} = 1.1-1.3). The profile description of a very regrated chernozem may be as follows:

$H_{e/k}$: slightly eluviated horizon of SOM accumulation: effervesces with 10% HCl in the lower portion; 30-45 cm thick; dark grey; amorphous SiO_2 in the upper portion of the horizon; plow layer is characterized by cloddy structure with powder; subplow layer – by cloddy structure with grains; earthworm casts; transition clear;

H_{pik}: upper transitive horizon; slightly illuviated carbonate-containing; 20-30 cm thick; well and uniformly supplied with SOM; dark ash-grey; cloddy or cloddy with nuts; slightly compacted; riddled with earth worm and mole tunnels; signs of R_2O_3 in the form of brownish tint; transition clear;

$P_{h(i)k}$: lower transitive horizon; very slightly illuviated; 30-40 cm thick; cloddy or prismatic-cloddy; slight R_2O_3 films on the faces of aggregates; abundant carbonates (veins); intensively transformed by earthworms; transition clear;

P_{hk}: the upper portion of parent material; 15-40 cm thick; soil organic matter is low content and nonuniformly distributed; dirty-brown; not hard clods; earthworm tunnels and "crotovinas"; carbonatic molds; transition clear, along a wavy line of former effervescence with 10% HCl;

P_k: parent material; dull-yellow with brownish tint; pseudomicellium and veins of carbonates.

Brownish podzolized chernozems encountered near the Carpathians, in Opillya, Rostochchya, Podilly, and the Prut-Dnister region have the profile: $H_{d\ 3-5}$ + $H_{e\ 35-50}$ + $H_{pi\ 30-40}$ + $P_{hi\ 25-30}$ + $P_{i\ 15-45}$ + P_k(P).

Wet (humid) podzolized chernozems in the Western Forest-Steppe zone have a profile: $H_{e(gl)\ 30-60}$ + $H_{pi(gl)\ 25-35}$ + $P_{hi(gl)\ 35-45}$ + $P_{i(gl)\ 15-25}$ + $P_{k(gl).}$

Gleyed from the surface podzolized chernozems occurring in the low-lying areas and saucer-like depressions "blyndtsya" (блюдця) among modal podzolized chernozems with watertable no less than 6 m from the surface have a profile: $H_{e(gl)\ 30-60}$ + $H_{pi(gl)\ 35-45}$+ $P_{higl\ 30-40}$ + $P_{(h)igl\ 20-40}$ + P_{gl}.

26.5.3. Podzolized Soils Composition and Properties

By texture, podzolized soils are approximately in equal measure represented by light, medium, and heavy loams. In northern and western Forest-Steppe the soils are coarse-silty or even sandy light loams. In the central part of the zone they are dominantly coarse-silty or silty medium loams, whereas in the south heavy loams dominate. With depth in the profile the texture becomes heavier, especially in the illuvial horizons (Table 74).

Table 74. Chemical Composition and Textural Features of Podzolized Soils (N.M.Breus, 1979)

Characteristics	Dark grey podzolized heavy loam in loess					Podzolized Chernozem heavy loam in loess				
	Horizons									
	H_e	H_i	I	P_i	P_k	H_e	H_{pi}	P_hI	$P_{(h)i}$	P_k
Layer, cm	10-20	40-50	70-80	110-120	130-140	10-20	50-60	80-90	110-120	130-140
Content of particles, %:										
< 0.01 mm	51.8	56.3	62.7	62.7	67.3	42.3	40.4	39.9	38.8	35.6
< 0.001 mm	30.7	32.7	39.8	38.4	45.8	25.3	24.8	24.4	23.2	22.6
Oxide content, %:										
SiO_2	79.1	77.5	77.0	75.7	73.6	80.3	80.7	80.8	80.8	81.2
Fe_2O_3	3.45	4.52	4.78	4.83	4.71	3.07	3.09	3.10	3.20	3.31
Al_2O_3	9.91	11.8	11.7	11.7	11.6	10.7	9.67	10.9	10.7	10.6
CaO	1.34	1.17	1.02	1.18	8.58	2.13	1.83	1.55	1.50	0.84
MgO	1.07	1.15	1.56	1.36	1.95	0.96	0.73	0.86	0.71	1.14
Na_2O	0.98	0.92	0.96	0.94	0.92	0.83	0.85	0.80	0.79	0.82
SiO_2/R_2O_3	9.60	8.30	8.30	8.10	8.30	10.8	11.8	10.7	10.8	10.85

There is not much eluvio-illuvial differentiation in the profiles of podzolized chernozems, but the contents of physical clay and colloids decreases with depth.

The analysis of the ratio SiO_2/R_2O_3 shows the podzolic process of soil formation to be more intensive in dark grey soils than in podzolized chernozems. Podzolized soils contain considerably more humus than grey forest soils (Table 75). In the former soils the humus content is moderate, whereas in the latter it is low or very low.

Table 75. Content and Composition of SOM in Podzolized Soils (D.I.Kovalishin and N.M.Breus, 1979)

Horizon & depth, cm	Total SOM content, %	Organic carbon content, %	C_{ha}	C_{fa}	C of insoluble residue	C_{ha} / C_{fa}
			% of organic C content			
Dark Grey Podzolized Heavy Loam Soil in Loess						
HE $_{10\text{-}20}$	4.29	2.49	34.7	33.9	31.4	1.02
HI $_{40\text{-}50}$	1.07	0.97	40.2	38.1	21.7	1.06
I $_{70\text{-}80}$	0.65	0.38	18.7	46.8	34.5	0.40
$P_{i\ 110\text{-}120}$	0.60	0.35	19.4	45.2	35.4	0.43
$P_{k\ 130\text{-}140}$	0.40	0.23	23.9	51.7	24.4	0.46
Podzolized Heavy Loam Chernozem in Loess						
$H_{e\ 10\text{-}20}$	5.19	3.01	40.9	36.2	22.9	1.13
$H_{pi\ 50\text{-}60}$	2.30	1.67	45.4	31.4	23.2	1.44
$P_hI_{80\text{-}90}$	0.96	0.56	30.4	37.7	31.9	0.81
$P_{(h)}I_{100\text{-}110}$	0.93	0.54	25.9	40.7	33.4	0.64
$P_{k\ 130\text{-}140}$	0.90	0.52	27.4	40.9	31.7	0.67

By the indices of exchangeable acidity podzolized soils are neutral or close to it. By hydrolytic acidity (3.89 for dark grey soil and 3.28 meq. per 100 g of soil for podzolized chernozem) the soils are slightly acid. Calcium is a dominant exchangeable cation (22-24 meq/ 100 g). Ca/Mg ratio changes in the range of 2.7-7.6 (Table 76).

Table 76. Physico-Chemical Characteristics of Podzolized Soils (O.M.Alexandrova and N.M.Breus, 1979)

Characteristics	Dark grey podzolized heavy loam soil in loess					Heavy loam podzolized chernozem in loess		
	Horizons							
	HE	HI	I	P_i	H_e	H_{pi}	P_{hi}	$P_{(h)i}$
Soil layer, cm	10-20	40-50	70-80	110-120	10-20	50-60	80-90	100-110
pH $_{KCl}$	6.3	5.5	5.8	5.8	6.4	6.5	6.5	6.7
Exchangeable cations, meq. per 100 g of soil:								
Ca^{2+}	23.6	20.4	17.5	22.8	22.5	17.0	13.7	16.3
Mg^{2+}	6.60	3.40	6.60	3.0	4.43	3.88	4.91	3.34
Na^{+}	0.18	0.22	0.23	0.17	0.31	0.26	0.26	0.10
K^{+}	0.09	0.74	0.61	0.61	0.41	0.51	0.51	0.51
Sum of exchangeable bases	31.5	24.8	24.9	26.6	27.6	21.6	19.4	20.2
Hydrolitic acidity, meq. per 100 g of soil	3.89	2.91	1.82	1.82	3.28	1.72	1.06	0.72
CEC, meq. per 100 g of soil	35.4	27.7	26.7	28.4	30.9	23.4	20.4	21.0
PBS, %	89	90	93	94	89	92	95	96

By physico-chemical characteristics, the podzolic process of soil formation is more intensive in dark grey podzolized soil compared to podzolized chernozem. But the two soils are close to each other in this respect. Agrochemical characteristics of podzolized soils are given in Table 77.

Table 77. Agrochemical Characteristics of Podzolized Soils (B.S.Nosko and S.P.Latyshev, 1994)

Characteristics	Dark grey podzolized heavy loam soil in loess			Heavy loam podzolized chernozem in loess		
	Horizons					
	HE	HI	P_i	HE	H_pI	$P_{(h)i}$
Soil layer, cm	2-12	40-50	110-120	0-10	50-60	100-120
Total content, % of:						
N	0.14	0.06	0.03	0.21	0.14	0.09
P_2O_5	0.12	0.09	0.09	0.13	0.12	0.11
K_2O	1.88	1.91	2.19	2.07	2.26	2.34
Content of available forms in Kirsanov extract (ppm):						
P_2O_5	95.8	54.7	Not determined	112	63.4	Not determined
K_2O	112	83.5		126	92.1	

Physical properties and properties related to soil water of podzolized soils are better than those of grey forest soils (Table 78).

Bulk density of dark grey podzolized soil in the surface horizon is high (1.30 g/cm^3), but in horizons below it, it's even higher (1.37-1.52 g/cm^3). Podzolized chernozem has this characteristic in values more favorable for crops. In the HE horizon it is within 1.22-1.23 g/cm^3, and the rest of the horizons are quite loose (1.14-1.18 g/cm^3).

Table 78. Physical Properties of Podzolized Soils (M.G.Iowenko, 1960)

Characteristics	Dark grey podzolized heavy loam soil in loess					Podzolized heavy loam Chernozem in loess				
	Horizons									
	H_e	H_i	I	P_i	P_k	H_e	H_{pi}	P_hI	$P_{(h)i}$	P_k
Soil layer, cm	0-10	30-40	60-70	90-100	120-130	0-10	30-40	60-70	90-100	110-120
Particle density, g/cm^3	2.63	2.70	2.71	2.72	2.72	2.68	2.70	2.70	2.70	2.70
1.38	1.30		1.41	1.37	1.52	1.22	1.23	1.18	1.14	1.16
Total porosity, % V	51	49	48	50	44	54	54	56	58	57
PWP, %m	8.3	9.7	11.8	11.6	10.0	11.3	11.2	11.1	13.6	12.3
FC, % m	24.5	22.6	25.5	23.4	23.2	26.2	26.3	26.4	26.8	27.4
Available water at FC, % m	16.2	12.9	13.7	11.8	13.2	14.9	15.1	15.3	13.2	15.1

All horizons in podzolized chernozem have satisfactory total porosity, whereas in dark grey podzolized soil these characteristics may be less than satisfactory.

Potential productivity evaluation by 100-point scale (A.I. Siry, 1998) showed it was within 40-55 points for dark grey podzolized soils and within 45-70 points for podzolized chernozems.

26.5.4. Agricultural Use and Practice Improvement of Podzolized Soils Improvement

Dark grey podzolized soils and podzolized chernozems are united into a single agroproduction group. Under conditions of erosion hazard the fields should be planned along the contours. According to ecologo-technologic grouping of lands, the practices of soil tillage, fertilizer application, and plant protection should be differentiated. On the slopes within 3-5° in steepness, row crops should not be employed in crop rotations, which (the latter) should consist of small grain crops and perennial grasses. Among the crops suitable for production on podzolized soils are winter wheat, sugar beet, corn, potatoes, alfalfa, annual grasses, etc. The soils are the best for fruit trees and berries.

Slightly acid reaction of podzolized soils demands regular liming. On loamy sands and sandy loams the rates of liming should be only a half of a complete rate computed by hydrolytic acidity value. On heavier soils, complete rates will not be too big, provided P_2O_5 and Zn are not bound in insoluble compounds.

Farm manure should be applied at rates of 14-16 mt per hectare of a crop rotation per year. Composts are a good supplement for FM.

26.6. Composition, Properties and Use of Regraded Soils

Regraded soils are the result of calcium carbonate formation in the formerly leached (and illuviated) horizons of podzolized soils, or even in the entire soil profile. But more often than not effervescence with 10% HCl is observed in the illuvial and illuviated horizons. Morphological evidence of carbonates often acquires the form of veins, molds, pseudomicellium, and even dolls (concretions –cutanes).

By texture regraded soils are much the same as their leached podzolized counterparts, and geographically are distributed in the same way. A slight redistribution of clay by the eluvio-illuvial type may be observed in regraded dark grey soils, whereas in podzolized chernozems it is practically nondetectable. The chemical composition of dark grey soils and chernozems is practically the same and does not change much down the profile. SiO_2 / R_2O_3 ratio is stable in both soils (10-11). That gives reason to believe that the sod-forming process of soil formation is more intensive in regraded soils compared with their leached counterparts. Humus content is approximately the same in dark grey and chernozemic regrated soils: 3.48

and 3.53%. Lighter-textured soils contain it less than heavier-textured ones. Humus type is fulvate-humatic in the upper and humate-fulvatic in the lower horizons of soil profile. Calcium dominates among the exchangeable cations. Percentage base saturation is over 90% and increases down the profile. Generally, physico-chemical properties of regraded soils are better than those of their leached analogues and make them more like typical chernozems.

Regradation is accompanied by the improvement of physical and soil water related properties of soils. They may have higher total and aeration porosity than the corresponding leached soils. By the content of available water at FC there is not much difference between these soils.

Potential productivity evaluation by the 100-point scale (A.I.Siry, 1998) showed it to be for dark grey regrated soils within 54-66 points and for regrated podzolized chernozems – within 69-77 points. Regrated soils are better supplied with available nutrients (from moderate to high gradations).

26.7. Forest-Steppe Chernozems and their Classification

Typical chernozems (modal) and typical chernozems humid are the most widespread subtypes, whereas leached chernozems occupy small areas. Classification of the Forest-Steppe chernozems is given in Table 79.

Table 79. Classification of the Forest-Steppe Chernozems (Research Institute of Soil Science and Agrochemistry, UAAS)

Subtype	Genus	Species
Leached humid Typical humid Leached Typical	Modal, carbonatic, eroded, saline from irrigation, meadowish from irrigation	Modal, by the extent of erosion, salinity, sodicity, the depth of the layer of SOM accumulation, humus content, depth of the upper saline horizon, depth of carbonate accumulation (effervescence) and ground water table

Carbonatic soil units are characterized by the distinct effervescence with 10% HCl from the surface. Carbonatic species are not present among podzolized and leached chernozems.

Deeply effervescing detectable carbonates lie deeper than in modal genuses. Such genus is encountered among typical chernozems.

Residually-carbonatic are formed in different carbonate materials like chalk, limestone, marl, etc. Solid particles of such rocks are encountered in soil profile. The H and HP horizons are underlain by the rubble of a solid rock eluvium. Usually effervesce from the surface.

Noncarbonatic effervescence and morphological carbonate features cannot be detected; they occur among typical, podzolized, and leached chernozems.

Eroded genuses differ from modal by partial or complete absence of the upper horizons (H, HP_k, etc) as a result of erosion by water or wind.

Washed on genuses are covered by the products of erosion from the upper-lying areas.

Soil species are identified by the depth of humus profile (H + HP): undeveloped (< 25 cm), short-profiled (25-45 cm), not deep (45-65 cm), moderately deep (65-85 cm), deep (85-120 cm), and extra deep (> 120 cm).

By the SOM content, the soils may be slightly humusified (< 3%), lowly humusified (3-6%), and moderately humusified (> 65).

By the extent of soil erosion, the following species are identified:

1) slightly eroded: up to ½ of the H horizon lost as well as up to 20% of the initial stock of SOM;
2) moderately eroded: from ½ to the whole of the H horizon lost as well as 20-40% of the initial stock of SOM; and
3) severely eroded: the profile is bereaved of the H, Hp, and partly Ph horizons. 40 to 60% of the initial stock of SOM is lost.

Leached chernozems divide into species according to the extent (depth) of leaching:

1) slightly leached: the line of effervescence is not deeper than 20 cm from the lower boundary of the upper transitive horizon (HP);
2) moderately leached: the line of effervescence is at the depth of 20 to 50 cm below the lower boundary of the upper transitive horizons (HP);
3) strongly leached: the above mentioned line lies over 50 cm deeper than the lower boundary of the HP horizon.

Lithologic series is identified by the character of parent material. Loess and loess-like loams are the dominant materials. There may be virgin, reclaimed, cultured, and irrigated variants of chernozems. Soil varieties are identified by texture.

26.8. Morphological Features of the Forest-Steppe Chernozems

Leached chernozems occur in small areas all over the Forest-Steppe zone, but their main territory are concentrated in the central and western Forest-Steppe as well as on the loess islands of Polissya. In the left-bank Forest-Steppe zone, at the boundaries with Polissya, they are found on high-level loess terraces among typical chernozems on well-drained areas.

By morphological features leached chernozems take an intermediate place between podzolized and typical chernozems. There is no eluvio-illuvial differentiation in their profile, and they are leached from the carbonates to the depth below $H + H_p$:

H: humus horizon 35-40 cm thick, dark grey, cloddy with powder in a plow layer, grainy with clods in a subplow layer; compacted; transition gradual;

$H_{p(i)}$: upper transition horizon 30-45 cm thick, grey with brownish tint; nutty-prismatic; compacted; mole tunnels; transition clear, wavy;

P_h: lower transition horizon; from 20-30 to 60 cm thick, slightly and nonuniformly “humusified”, spotty; mole tunnels; transition distinct (sharp), wavy with tongues;

P_k: parent material, straw-colored loess or loess-like loam with carbonate efflorescence and veins. Humid subtypes of leached chernozems in the western Forest-Steppe zone have the horizons: $H_{40\text{-}50} + H_{p(i)\ 50\text{-}60} + P_{h(gl)\ 30\text{-}40} + P_{kgl}$.

Typical chernozems are the most widespread subtype in the forest-Steppe zone. Their main areas are located in the central and the eastern Forest-Steppe parts on loess plateaus and their slopes, as well as on high-level loess terraces with a low ground water table. Such soils were formed under “stepped” meadows and meadow steppes in loess. The profiles are characterized by most typical chernozemic features:

$H/_k$: horizon of SOM accumulation, reaching the depth of 45-55 cm; dark grey, powdery with clods in a plow layer; grainy with clods in a subplow layer, loose, friable or only slightly compacted; carbonatic in a lower portion; rare "crotovinas" and earthworms tunnels and casts; transition gradual;

H_{pk}: upper transition horizon; 35-45 cm thick, slightly less rich in SOM than H horizon; carbonatic, dark grey with almost undetectable brownish tint; cloddy; slightly compacted; earthworm and mole tunnels; transition gradual;

P_{hk}: next transition horizon; 30-40 cm thick; grey with brown tint; spots of mole tunnels; cloddy with friable clods; slightly compacted; pseudomicellium of lime; transition gradual.

$P_{(h)k}$: a spotty horizon of abundant mole tunnels; 30-80 cm thick; dirty-yellow, mottly, abundant carbonate pseudomicellium, cloddy, loose, transition clear;

P_k: parent material, straw colored loess with thin veins of $CaCO_3$.

Cultivated soils have their humus content within 3-5%, whereas virgin variants may have it within 5-9%. Humus type: humatic or fulvate-humatic; the structure of virgin variants is well expressed with abundant buckwheat kernels of grainy aggregates.

Humid subtypes of typical chernozems occupy the Western regions. Their profile may have the following features: $H/k + H_{pk} + P_{hk} + P_{hkgl} + P_{kgl}$. The humidity of climate is the main reason of their formation. The layers of loess material may be underlain by limestones and marls, impenetrable to water.

26.9. Forest-Steppe Chernozems Composition and Properties

Typical chernozems are most widespread mapping units in the Forest-Steppe zone. They are characterized by considerable SOM and nutrient element accumulation in the upper 0.5 m layer of the soil profile. All agents of soil fertility are well balanced in such soils. By texture, medium and heavy loams are dominant (40.4 and 34.5%, respectively); light loams occupy 25% of their area, 0.5% remains for loamy sands and light clays. The texture becomes heavier from the north to the south. Physical clay fraction gradually decreases with depth (Table 80).

Table 80. Chemical Composition and Textural Features of Typical Chernozems in Loess (N.M.Breus, 1979)

Characteristics	Typical deep-profiled low-humus medium loam chernozem					Typical deep-profiled moderate-humus heavy loam chernozem				
	Horizons									
	H_a	$H/_k$	H_{pk}	P_{hk}	P_k	$H/_{ka}$	$H/_k$	H_{pk}	P_{hk}	P_k
Soil layer, cm	0-10	25-35	45-55	110-120	210-220	10-20	30-40	60-70	80-90	180-190
% of particles:										
< 0.01 mm	42.2	37.3	34.4	30.8	30.7	52.1	53.1	50.7	50.1	52.3
< 0.001 mm	24.9	22.1	21.0	19.3	19.9	33.9	33.7	33.6	32.4	33.1
Oxide content, %:										
SiO_2	78.2	75.3	75.4	73.8	66.8	67.9	Not determined	68.5	71.3	62.7
Fe_2O_3	3.24	3.34	3.47	3.04	3.32	3.75		3.70	3.85	4.30
Al_2O_3	9.35	9.06	8.58	9.11	7.35	12.8		12.6	13.7	13.1
CaO	2.19	3.24	5.56	7.39	6.72	2.07		2.51	1.86	6.84
MgO	1.02	1.06	1.16	1.62	1.44	0.97		0.92	1.06	1.47
$SiO_2 : R_2O_3$	11.6	11.4	11.9	11.4	11.4	7.66	>->	7.76	7.53	6.75

Typical chernozems with "slight" humus content are found in the northern part of the zone. Low humus content chernozems are dominating in the zone, prevailing in the Central Forest-Steppe. Chernozems with a moderate (or average) humus content occur in the southern and eastern regions. The stocks of soil humus in the humus-containing part of the profile in light loams are within 260-360 mt/ha. Medium loams: 390-480 mt/ha, and heavy loams and light clays: 460-560 mt/ha (N.M.Breus, 1988). Humic acid dominate over fulvic ones (C_{ha} / C_{fa} = 1.53-3.41).

Calcium humates is the most typical organic-mineral derivative of typical chernozems (Table 81).

Table 81. Soil Organic Matter in Typical Chernozems in Loess (N.M.Breus, 1979)

Horizon and depth, cm	Total SOM content, %	Total organic carbon content, %	C_{ha}	C_{fa}	C residual	C_{ha} / C_{fa}
			% of total organic carbon			
Medium Loam Deep Typical Chernozem With Low SOM Content						
$H_{a\ 0-10}$	4.58	2.66	44.2	22.4	33.4	1.97
$H/_{k\ 25-35}$	4.26	2.47	47.8	20.6	31.6	2.32
$H_{pk\ 45-55}$	3.55	2.06	47.2	18.2	34.6	2.59
$P_{hk\ 110-120}$	1.63	0.94	29.6	31.0	39.4	0.95
Heavy Loam Deep Typical Chernozem with Moderate SOM Content						
$H/_{k\ a\ 10-20}$	5.61	3.25	36.2	23.6	40.2	1.53
$H/_{k\ 30-40}$	4.67	1.97	40.7	20.7	38.6	1.97
$H_{pk\ 60-70}$	3.48	2.01	47.4	13.9	38.7	3.41
$P_{hk\ 80-90}$	2.29	1.33	32.6	12.8	54.6	2.54
$P_{k\ 180-190}$	1.74	0.89	23.1	25.7	51.2	0.90

pH $_{H2O}$ in a plow layer is about 6.8-7.3; hydrolytic acidity 0.2-0.7 meq/100 g. In deeply effervescent soils these characteristics are 5.9-6.0 and 2.3-2.6, respectively (N.M.Breus, 1981). CEC depends on soil texture and SOM content. Medium loam soil's CEC reached 37.0 meq/100 g (18.7 in parent material). In heavy loam soil, CEC was about 40 meq/100 g (29,1 in parent material). Calcium dominates in SAC, exceeding magnesium 5-6 times (Table 82).

Table 82. Physico-Chemical Characteristics of Typical Chernozems in Loess (N.M. Breus, 1979)

Characte-ristics	Medium Loam Typical Chernozem in Loess					Heavy Loam Typical Chernozem in Loess				
	Horizons									
	H_a	$H/_k$	H_{pk}	P_{hk}	P_k	$H/_{ka}$	$H/_k$	H_{pk}	P_{hk}	P_k
Soil layer, cm	0-10	25-35	45-55	110-120	210-220	10-20	30-40	60-70	80-90	180-190
pH $_{H2O}$	6.8	7.4	7.4	7.9	8.3	6.7	6.8	7.0	7.2	7.3
Exchangeable cations, meq/100 g: Ca^{2+} Mg^{2+} Na^{+} K^{+}	 29.9 3.61 0.33 0.42	Not determined				 35.3 4.13 0.13 0.41	Not determined			
Hydrolytic acidity, meq/100 g:	0.70	Not determined					Not determined			
CEC, meq/100 g	37.0	32.8	32.2	26.0	18.7	40.0	40.8	35.7	33.7	29.1
% $CaCO_3$	–	0.38	3.54	8.00	7.09	0.97	3.68	4.46	3.31	12.2

Typical chernozems are well provided with total and available nutrients (Table 83).

Table 83. Agrochemical Characteristics of Typical Chernozems in Loess (N.M. Breus, 1979, A.I. Siry, 1985)

Characteristics	Medium Loam Typical Chernozem in Loess					Heavy Loam Typical Chernozem in Loess				
	Horizons									
	$H/_{ka}$	$H/_{k}$	H_{pk}	P_{hk}	P_{k}	$H/_{ka}$	$H/_{k}$	H_{pk}	P_{hk}	P_{k}
Layer, cm	0-10	25-35	45-55	110-120	210-220	10-20	30-40	60-70	80-90	180-190
Total content,%:										
N	0.30	0.24	0.20	0.10	0.02	0.29	0.26	0.21	0.19	0.011
P_2O_5	0.14	0.13	0.11	0.09	0.07	0.15	0.15	0.15	0.11	0.03
K_2O	0.21	0.19	0.19	0.20	0.20	0.24	0.24	0.23	0.21	0.19
Content of available forms in Chirikov extract (ppm):										
P_2O_5	185	121	Not determined			196	145	Not determined		
K_2O	171	102				183	117			

Physical and soil-water related properties of typical chernozems are good in most cases. It pertains to the bulk density and total porosity. At field capacity (FC) the soils are well provided with available water (Table 84). It is necessary to take care of soil aeration porosity at FC, which should not be below 15% of the total volume. But, so far, in the majority of cases, capillary porosity is within 60-65%, of the total and noncapillary (aeration) – 35-40%, which makes good aeration.

Table 84. Physical and Soil-Water Related Properties of Typical Chernozems in Loess (A Source-Book on Agrohydrologic Properties of Ukrainian Soils, 1965)

Characteristics	Medium Loam Typical Chernozem in Loess							Heavy Loam Typical Chernozem in Loess
	Horizons							
	H_a	$H/_k$	H_{pk}	P_{hk}	$H/_{ka}$	$H/_k$	H_{pk}	P_{hk}
Layer, cm	0-10	30-40	50-60	110-120	10-20	30-40	60-70	80-90
Particle density, g/cm^3	2.55	2.62	2.62	2.63	2.62	2.64	2.68	2.70
Bulk density, g/cm^3	1.23	1.14	1.13	1.17	1.13	1.18	1.44	1.46
Total porosity, % V	51.8	56.5	56.9	55.5	57.0	55.3	46.3	46.0
PWP, %m	9.88	10.9	11.3	8.94	16.9	18.6	18.2	17.9
FC, % m	26.7	23.9	22.7	20.0	39.0	32.6	29.2	27.2
Available water at FC, %	21.3	16.3	15.4	17.3	25.0	16.5	15.9	13.6

Leached chernozems do not considerably differ from typical ones by their natural productivity. They are in the same agroproduction group. They show effervescence with 10% HCl at the depth of 80-100 cm. The potential productivity of typical chernozems, according to A.I.Siry (1988), is from 60 to 90 points on 100-point scale; the heavier the texture, the higher the grade point. Leached chernozems' grades are within 63-75 points.

26.10. Meadow-Chernozemic Soils of the Forest-Steppe Zone

Meadow-Chernozemic soils of the Forest-Steppe zone are semihydromorphic analogues of chernozems. They form under meadow-steppe grasses with diverse associations of species. One of their formation conditions is an increased wetness from atmospheric precipitation, or ground water, or both. Ussually, ground water table is within 2-4 m from the surface. The soils locate mainly in the saucer-like depressions of mesorelief. They have the signs of hydromorphism in the soil profile lower portion, like olive, dull-grey and rusty spots as well as marly components of the parent material. Meadow chernozemic soils are used as farmland (671 thousand ha) and plowland (530 thousand ha). According to their classification, four subtypes are identified in the Forest-Steppe zone: modal, podzolized, solonetzic (sodic), and solodized (*solod* or *solot* is a Russian term used also in Canada).

There are the following genuses of meadow-chernozemic soils:

1) **modal** (ordinary, typical), as the name implies; the word modal is not used in the soil name;

2) **carbonatic,** as the name implies, the soil profile shows effervescence with 10% HCl;

3) **podzolized:** with eluvio-illuvial profile differentiation;

4) **sodic** (sodicity-affected): with eluvio-illuvial differentiation caused by exchangeable sodium (solonetz-like);

5) **saline:** with soluble salts in the soil profile identified morphologically or detected in the laboratory by soil extract analysis;

6) **washed on:** being low-lying, the soil suffers the inflow of eroded material from surrounding upland areas. Soil tillage also favors the moving of the soil into depressions of topography.

Meadow-chernozemic soils are divided into species by the humus profile depth and the humus content in the surface horizon (plow layer) in the same ways as chernozems.

By morphological features, these soils are similar to chernozems, having, at the same time, some specific signs. Their humus content and the depth of humus accumulation may be greater. They have the signs of gleying which may disappear in dry years and appear again in wet ones. Below is the description of one of the profiles:

H: horizon of humus accumulation; 35-55 cm thick; dark grey, plow layer powdery with clods; subplow – grainy with clods and nuts; transition gradual;

$H_{p/k}$: upper transition horizon; 20-30 cm deep, dark grey with brownish tint; grainy or grainy with clods and nuts; slightly compacted; effervesces in the lower portion; transition gradual;

$P_{hk(gl)}$: lower transition horizon; 15-30 cm thick; grey brown; cloddy or cloddy with nuts; compacted; carbonates as “mold’; brownish-rusty spots of gleying; transition clear;

$P_{kg}l$: parent material; loess; brownish-yellow; olive and dull grey spots; veins of $CaCO_3$; becomes more “marly” with depth.

Podzolized meadow-chernozemic soils may have the horizons: He + H_{pi} + P_{hi} + +$P_{i(h)gl}$ + P_{kgl}.

Solodized meadow-chernozemic soils prevail in locked saucer-like depressions in loess. They may stay flooded (puddled) periodically. The soil profile is believed to be formerly saline to a certain extent. As a rule, soluble salts are leached below the solum.

The soil profile may have the following features:

$H_{e(gl)}$: eluviated (solodized) horizon of SOM accumulation: 25-45 cm thick; cloddy with grains; amorphous silica on the faces of soil aggregates; transition clear;

H_{pigl}: upper transitive horizon; 20-35 cm thick; grey with dull grey shade; cloddy or cloddy with nuts; may be prismatic; compacted; rusty spots on the faces of aggregates, transition clear;

P_{hgl}: lower transitive horizon; 20-40 cm thick; nonunifirmly colored; mottly; with the dominance of dirty-olive and dull grey spots; mole tunnels; cloddy-prismatic; compacted;

P_{kgl}: loess-like gleyed parent material, which becomes more marl-like with depth;

Among the meadow-chernozemic soils there are encountered saline ones from the surface or from a certain depth. The former are characterized by the following sequence of horizons: $H_{ssl(k)\ 15\text{-}20}$ + $H_{pks(i)\ 15\text{-}30}$ + $P_{hkgls\ 30\text{-}60}$ + P_{kgl}, and the latter – by the following one: $H_{e\ 35\text{-}40}$ + $H_{pi\ 25\text{-}40}$ + $P_{hkgls\ 20\text{-}40}$ + P_{ksg}.

Meadow-chernozemic soils may be productive, but their effective use demands some practices which depend on their genetic nature.

26.11. Agricultural Use and Improvement of the Forest-Steppe Chernozems

Typical and leached chernozems occupy large level areas and gentle slopes. They are most productive of the Forest-Steppe soils. Such soils are good for small grain, forage, and vegetable crops, as well as for the production of fruits and berries. They do not require serious amendment, but the practices of soil and water conservation and economic use are important. Farming on the contours is a necessity even for such soils. Combined (plow and non-plow) tillage is the best for most crops. Plowing should be used for row crops and be combined with the application of FM, which should be applied at rates of 12-15 mt per hectare of a crop rotation per year to "fight" a negative balance of SOM. Leached chernozems as well as deeply effervescent typical chernozems, especially with a low humus content may need a regular application of low rates of lime. Overliming may cause the binding of available phosphorus, zinc, and other nutrients in insoluble forms. Every metric ton of NPK fertilizers on such soils may need approximately 0.5 metric ton of $CaCO_3$. Solodized soils in depressions may need additional liming as well as slitting the soil with disks in the way of tillage.

Some depressions with stagnant water on the surface (puddling) should not be used for planting of winter and spring small grain crops and for orchards. Row crops should not be grown on the sloping lands over 3 degrees. To make farming more biological it is worthwhile to apply nitrogen fixation and phosphorus-mobilizing preparations.

Without restriction of tillage operations which should be timely and effective, as well as without perennial grasses in crop rotations, it is impossible to impede the processes of soil structure pulverization and resulting compaction. Non-plow tillage operations with mulching are effective in increasing earthworm populations and improving soil tilth.

Questions & Assignments

1. Show the boundaries of the Forest-Steppe zone on the soil map of Ukraine. What is its total area?
2. Which are the general features of the soil cover in the Forest-Steppe zone? What does the name Forest-Steppe imply?
3. Name the essential characteristics of climate in the Forest-Steppe zone.
4. Speak on the parent materials of soils.
5. Name the horizons of light grey and grey forest soils. How are their profiles different?
6. Speak on the morphological features of dark grey podzolized soils. How are they different from podzolized chernozems?
7. How and why podzolized chernozems are different from typical ones?
8. What are regarded (repeated by saturated) soils and how were they formed?
9. Speak on the morphological features of leached chernozems. Speak on the differences between podzolized and typical chernozems?
10. Speak on the agricultural use and the improvement of podzolized soils.
11. Which Forest-Steppe chernozems do you know? How and why do they differ from each other?
12. Which typical chernozems may need liming? How should lime be applied to them?
13. Speak on the physical properties of typical chernozems. What should aeration porosity be at FC in their plow layer?
14. Speak on the potential productivity of podzolized and chernozemic Forest-Steppe soils.

Chapter 27

Soils of the Steppe Zone of Ukraine

The Ukrainian Steppe zone extends to the south and south-east of the Forest-Steppe zone. Its length from the south-west to the north-east reaches 1100 kilometers and from the north to the south – up to 500 kilometers.

By the peculiarities of climate and soils, the Steppe zone is divided into the northern, central, and southern subzones. Completely or partly, it takes in Kirovograd, Odesa, Dnipropetrovsk, Zaporizhzhya, Kherson, Mykolaiv, Donetsk, Lugansk, and Kharkiv regions (oblasts), as well as the northern (plainland) part of the Autonomic Republic of Crimea. The zone's natural division into north and south is perfectly reflected by the character of soils, the north corresponding to ordinary and the south- to southern chernozems. The partly conventional division line between these really natural physico-geographic formations extends through Izmail, Artsyz, Tyraspol, Rozdilna, Berezivka, Shiroke, Nikopol, and Osypenko. The northern chernozemic Steppe zone extends to the north of this line, to the boundary with the Forest-Steppe, whereas the southern chernozemic Steppe stretches to the south up to the northern boundary of the Arid Steppe zone. On the Crimean peninsula, the southern Steppe comes again and extends to the south till it makes a transition to the mountain region. Another area of the southern Steppe is located in the very east of Ukraine (Lugansk oblast).

Including the Arid Steppe zone, with its nonchernozemic dark chestnut and chestnut soils, the area of Ukrainian Steppe is about 25 mln hectares (40% of the Ukrainian territory). 45% of Ukrainian plowland are in the Steppe. Slope lands (on the slopes exceeding 1° in steepness) occupy 47.4% of farmland. Eroded lands exceed 38% of the total area, which is 11 mln hectares, including 5 mln hectares of land which suffered from severe wind erosion.

The soil cover of the northern and the central Steppe is genetically homogeneous. There are 69 soil species, but ordinary chernozems with low or moderate SOM content are most widespread. Micelliary carbonatic and sodic subtypes occupy small areas, by the "diversity of erodedness", or, rather, xeromorphicity.

The southern Steppe is the region of southern and micelliary carbonatic chernozems, among which sodic or sodicity-affected chernozems may be found. The soil diversity in the southern Steppe is made up of 84 soil species, among which there are many eroded or xeromorphic soils with shortened profiles.

27.1. Factors of Soil Formation

Climate in the Steppe is moderately warm with insufficient humidity. Summer is hot, winter is mild, sometimes cold but with not much snow. In the west the climate is warm and relatively more humid than in the east, where it is hot and more continental which means "with more severe winters". On the Donetsky upland (ridge), as well as on the more western uplands, like Podilska, Prydniprovska, and Moldavska, the annual precipitation is 30-40 mm bigger than on the surrounding lower plainland. The most arid regions are the Prydniprovska lowland and Crimea.

Climate in the northern and the central Steppe is moderately warm and moderately continental. The mean monthly air temperature in January is within the range of 4–8°C, in July – 21–23°C. The period of crop growth and development lasts for 158-188 days. The annual sum of air temperatures over 10°C is in the range of 2800-3200°C. Mean annual rainfall (and snowfall) is within 425-500 mm. The greater part of rainfall (230-280 mm) comes in the warm season of the year. Rainstorms and downpours with 100-150 mm of rainfall in a day are quite frequent and operative in intensifying soil erosion processes. Snowmelt water may wet the soil to the depth of 2-3 meters and may not reach ground water.

Water evaporation from the soil surface considerably exceeds the annual precipitation. Humidity coefficient is, therefore, less than 1 (0.6-0.8). Meteorologic observations in the course of many years give reason to conclude that climatic conditions in the northern and central Steppe of Ukraine are relatively favorable for the growth of small grain crops, especially winter wheat. They are favorable for the production of sunflower, corn, hemp, vegetables, watermelons, (melons), fruits, rape, soybean, etc. But unstable water regime and periodic droughts, as well as water erosion and windstorms, make farming hazardous here.

Climate of the Southern Steppe is moderately droughty and averagely continental. It is characterized by a high air temperature and low rainfall. In January mean monthly air temperature is 2.5-3° below zero. In July it is 23-24° above zero. The period of crop growth and development lasts from 200 to 230 days. The sum of temperatures over 10°C is in the range of 2900-3600°C.

Mean annual precipitation is within 300-450 mm. Over 200 mm fall during the period of crop growth. Heavy rainfalls cause the soil erosion. In the southern Steppe subzone long periods without rains are observed every year. Once in every two years such periods exceed 40 days, which causes soil droughts. Eastern dry winds dominate here in July-August. But it is spring and autumn winds that cause dust storms, which damage and sometimes completely destroy seedlings and blow out the surface layer of the soil.

The southern Steppe is even more hazardous for farming than the northern and the central Steppe. Practices of water harvesting and irrigation are a necessity here. The main crops of the southern Steppe are winter wheat, sunflower, spring barley, and grapes. Irrigation creates favorable conditions for the growth of rice, corn, vegetables, forage crops and fruits. Sufficient amounts of heat and solar radiation favor the production of mellons, watermellons, hemp, curly-fibre flax, and ether-and-oil bearing crops.

Relief. By relief the Steppe zone is a plain land, but it is not everywhere similar in origin, geologic structure, and absolute altitudes. These factors cause a different intensity of erosion processes, as well as the meso- and microrelief landforms in the regions. The northern Steppe has a plain, moderately wavy land surface with numerous hollows and gullies, as well as river valleys 75-150 m deep. Near the rivers, the stream pattern becomes thicker and erosion processes – more intensive.

The central part of the Steppe zone lies within the Prychornomorska lowland which in many places is just a flat plain. Some uplands 150-300 m in altitude come on it from the north, north-east, and east. (Besarabska, Podilska, Prydniprovska, the edges of the Medium Russian upland, and the Donetsky ridge, respectively), which are characterized by well developed stream patterns and erosional landforms. Left banks of the rivers have much less developed landforms, compared with right ones, with their thick patterns of valleys and hollows, and steep banks, suffering from stream-bank erosion and land slides. Southern subzone is the most level and flat one, with slightly developed stream pattern. Meso and micro land forms are developed in the Prychernomorska lowland. The plainland here is covered by panlike depressions with gentle, hardly noticeable slopes. They may be up to 7-8 and more km wide. Their depth is not big. They are called “pods”. In spring they accumulate snowmelt water. Pods are the only water-collecting catchments in the southern Steppe. Some of such large collectors may remain filled with water till midsummer and later. As their bottoms are practically impenetrable, they lose their water by evaporation. They help to keep air humidity more favorable for plants, animals, and humans. Higher air humidity is operative in the prevention of dust storms, but if it is too hot and the air – too humid, the human beings have to suffer more.

Vegetation. Natural plant species in the steppe zone were formerly represented mainly by cereal grasses capable to withstand insufficient soil moisture and air humidity. Drought-resistant narrow-leaved species like feathergrass, needle grass, fescue wormwood, salvia, yellow alfalfa, bluegrass, tulips, etc, were most widespread in the virgin steppe. Now the “native” vegetation is preserved in reserves. Further to the south, as the soil salinity increases, a number of wormwoods and halophytes in natural associations increases. Ephemeral species, tulips, and special herb forms like thimble-weed, were popular in the virgin southern steppe. Meadow grasses and sedges form associations in the pods and other lowland areas with high watertable.

By the character of natural vegetation, the Ukrainian Steppe is distinctly divided into two subzones: northern – with meadow grasses, feather grass, and needle grass and southern – with feather grass and needle grass associations. From the north to the south, grass associations become thinner. On soils formed in non-loess parent materials on the Donetsky ridge, wormwood and petrophytes form thin associations.

Parent materials. Loess in the dominant parent material in the Steppe zone of Ukraine. Loess deposits cover watershed plateaus and terraces of the river valleys. By texture, loess in the Steppe is mainly heavy loam. Within the Prychernomorska lowland, the texture of the loess deposits is dominantly light clay. They are underlain by red-brown clays, which, in turn, are underlain by limestones. On the Pryazovska upland the deposits of loess may be medium clay. In the southward direction the loess, generally, becomes heavier and more saline.

The Donetsky highland is an elevated area reaching 367 m in absolute altitude. This highland is intensively eroded and has a thick pattern of streams. Loess dominates here as a parent material of soils, but in many places non-loess materials come to the surface, like sandstones, limestones, chalk, schists, red-brown clays, etc., the eluvium and deluvium of which serve as parent materials of soils. The soils formed in them are skeletal and contain rubble.

Alluvial deposits of recent and ancient origin are parent materials in floodplains and on the terraces. They are mainly light-textured. The factors of the soil formation in the Steppe are combined in such a way that chernozems with maximum SOM (humus) content occupy the band between the southern Forest-Steppe and the northern Steppe.

27.2. Soil Genesis in the Steppe Zone

Soil formation in the Steppe zone of Ukraine occurs under conditions of a moderately warm climate with an insufficient moisture regime, level topography, under grass-like, mainly cereal vegetation, in carbonate-containing parent materials (loess and loess-like loams).

The leading process of soil formation in the Steppe zone is the accumulation of SOM (humus) which causes the formation of humuso-accumulative horizons with an increased content of nutrient elements. Root systems of steppe grasses by weight exceed the biomass on the surface of the soil 4-20 times, and steppe wool (H_0 horizon) and plant residues on the surface exceed green biomass of plants from 2 to 5 times. The annual amount of plant residues coming to the soil in the virgin steppe is within 15-20 mt/ha.

The root mass of cereal perennial grasses renews up to 65% of its weight. It dies out later than the surface part. This prevents the complete mineralization. Processes of organic matter accumulation and its composition in the Steppe soils have a number of specific features:

1) root and surface residues are rich in ashy and nutrient elements;

2) steppe “wool” layer has a buffering effect on the organic residue decomposition preventing its complete mineralization and stimulating humification;

3) organic residues do not enter the soil, but remain in the soil; it is practically impossible to separate small roots from the soil;

4) organic matter mineralization coincides in time with intensive growth and development of plants, supplying the latter with nutrient elements;

5) organic matter enters the soil continuously with maximum accumulation close to the surface.

Humic acids dominate in umus of the steppe chernozems. Fulvic acids here have a higher molecular mass and are more complex, compared to the soils of Polissya. The presence of Ca^{2+} in soil solution $[Ca(HCO_3)_2]$ favors the formation of calcium humates, which, in their turn, prevent humus from leaching down the profile and favor the formation of water-proof grain aggregates.

Ordinary and southern chernozems are the most widespread soils in the Steppe, occupying over 90% of its area. Salt accumulation, which increases with aridity, is one of the features of soil formation. Micelliar-carbonatic chernozems (ordinary and southern) are less wide spread. The total area of ordinary chernozems is about 12, 251 thousand hectares, and that of southern chernozems – about 10, 713 thousand hectares. The rest of the zone is occupied by chernozems in non-loess parent materials, meadow-chernozemic, meadow, and other soils. Salt-affected soils of the Steppe zone are represented by saline, sodic, and solodized mapping units.

27.3. Chernozems of the Steppe Zone and their Classification

Ordinary chernozems are widespread in northern and central parts of the Ukrainian Steppe zone. In the south, the Chernozemic steppe reaches the outcrops of the Crimean mountains. The Crimean province of southern chernozems borders in the north with the Arid Steppe zone. According to the soil classification (Table 85), there are four subtypes of ordinary chernozems: modal (typical by features), micelliar-carbontic, sodicity-affected, and secondarily (as a result of irrigation) sodicity-affected from the surface. Southern chernozems also have four subtypes: modal, micelliar-carbontic, sodicity-affected, and secondarily sodicity-affected from the surface.

Table 85. Classification of the Steppe Zone Chernozems (Research Institute of Soil Science and Agrochemistry, 1988)

Subtype	Genus	Species
Ordinary: ordinary micelliar-carbonatic; ordinary sodicity-affected; ordinary secondarily sodicity-affected from the surface **Southern:** southern micellian-carbonatic; southern sodicity-affected; southern secondarily sodicity-affected from the surface	Modal, carbonatic, eroded, blown out, washed-on, deeply plowed, with secondary salinity, with secondary sodicity, with secondary eadowness; salinity affected with secondary meadowness, mocharic	Modal, by the extent of erosion, salinity, sodicity, depth of SOM accumulation $(H+H_p)$, humuscontent, depth of the first from the surface saline horizon, depth of effervescence and watertable level
	Soil division into lithologic series, variants, and varieties is the same as for the typical chernozems	

Division into species is made similar to typical chernozems. But specially identified are secondarily (as a result of irrigation) saline, sodic soils, as well as meadowish and secondarily meadowish ones. Morphological features, described in the field, must be complimented by laboratory analyses, to determine the species.

27.4. Morphological Features and Soil Profile Description

Ordinary chernozems lie on watershed plateaus, their slopes, and loess terraces of the river valleys, where they formed under feather grass – meadow grass associations mainly in loess and loess-like loams, as well as in red-brown clays and other nonloess parent materials. Ordinary chernozems formed in loess have in their profiles a characteristic form of carbonate accumulation – pea-like concretions (called *bilozirka* in Ukrainian: bilozirka = white star), lying below the layer of SOM accumulation. Chernozems formed in limestone eluvium and other nonloess materials do not have *bilozirka*. Transition horizons may have the other forms of carbonate accumulation, like pseudomiccelium or veins. From the north to the south, SOM content decreases from 6 to 3%. The depth of SOM accumulation also decreases. The horizons of carbonate, gypsum, and soluble salt accumulation come closer to the surface. In the northern part of the zone, ordinary chernozems contain gypsum and soluble salts below 4 m of depth, whereas in the south salt accumulations are detected at a depth of 2-2.5 m.

Ordinary deep chernozems with moderate SOM content occur in the northern part of the zone, along the boundary with the Forest-Steppe zone and on the Donetsky ridge – on the watersheds and their gentle slopes. They effervesce from the depth of 40-70 cm. The depth of SOM accumulation exceeds 85 cm.

Below goes follows a brief soil profile description:

H: surface horizon of SOM accumulation 40-50 cm thick; uniformly impregnated with humus; dark-grey; plow layer powdery with clods, subplow – grainy with clods; many earthworm casts; transition gradual.

$H_{p/k}$: upper transition horizon; 25-35 cm thick; dark grey with a slight brownish tint; carbonate molds in the lower portion; grainy or grainy with clods; porous; transition gradual;

P_{hk}: lower transition horizon; 30-40 cm thick; dark brown or greish brown with abundant carbonatic molds on the faces of aggregates; earthworm channels; holes from the root penetration; transition clear;

P_k: loess with mole tunnels and numerous carbonate concretions (*bilozirka*).

Sodicity-affected ordinary chernozems were formed mainly in saline clays of watershed plateaus or on loess terraces. Though exchangeable sodium does not ordinarily exceed 5% (and, occasionally, 10%) of the CEC, the narrow Ca/Mg ratio favors the formation of morphologic features peculiar to sodicity-affected soils. Plow layer is pulverized. Transition horizons are more than ordinarily compacted.

In the last decades, considerable areas of ordinary chernozems suffered secondary (caused by irrigation) sodicity, as they were irrigated with poor-quality water in which equivalent concentration of Mg^{2+} exceeded that of Ca^{2+}, or Na^{+} was in excess of $(Ca^{2+}+Mg^{2+})$, for a long period of time. The dispersion of soil colloids makes a wet soil sticky (adhesive) and a dry soil – very hard and with cracks. The upper portion of soil profile is differentiated into eluvial and illuvial horizons, though not very distinctly. Moderately sodicity-affected ordinary chernozem may have a profile consisting of HE (H_e) horizon, 25-30 cm deep, darkish grey with distinct amorphous silica visible on drying; blocky or cloddy; upper transition horizon (H_{pi}), 15-20 cm thick, dark grey, sometimes with brownish tint, compacted, nutty or prismatic, with colloidal films on the faces of prisms; lower transition horizon (P_{hik}), 20-30 cm thick, dark brown, cloddy-prismatic, and parent material (P_k) – straw-colored, with brownish tint.

Ordinary chernozems, secondarily sodicity-affected from the surface, are encountered on irrigated land areas and have the signs on sodicity mainly in a plow layer. In early spring or late autumn the surface acquires a whitish tint, caused by skeletal quartz washed from humus and other colloidal coatings. When wet, the soil is sticky. It is prone to puddling and crust formation. Plowing results in large blocks on the surface of the soil.

Micellian-Carbonatic Ordinary Chernozems are widespread on the territory of Prydunayska (near the Danube) soil province, lying between the Danube and the Dnister. A peculiar feature of such soils is carbonate accumulation in the form of pseudomicellium alongside “bilozirka”. The formation of abundant carbonate mold is caused by a specific hydrothermal soil regime, which, in its turn, increases the mobility of soil carbonates. The climate on a given territory is milder than elsewhere in the zone: mild winter and warm summer with sufficient rainfall.

The soils are highly biogenic: with numerous earthworm casts, mole channels, and a greater depth of the layer of SOM accumulation, better developed soil structure, and great total porosity. Such soil features caused the leaching of soluble salts to a greater depth.

Humus content in the plow layer is within 4-6%. Effervescence is detected in a subplow layer. Horizon of SOM accumulation (H) is 40-60 cm deep; the subplow layer is sometimes with $CaCO_3$ mold on the faces of aggregates. The upper transition horizon (H_{pk}) is 20-30 cm thick, dark grey with brownish or greish tint, and abundant carbonate needles giving the visual impression of mold. Lower transition horizon (PH_k) is 20-30 cm thick, cloddy with nuts and abundant relic earthworm casts. The Phk horizon is also 20-30 cm thick, rich in nuts and earthworm pellets of elliptical outline; abundant carbonate molds and occasionally *bilozirka* are encountered in the lower portion. But the greater part of *bilozirka* is found in the parent material.

Southern chernozems were formed in the southern subzone's hot climate under feathergrass, needlesgrass, fescue, and wormwood steppe associations. As the climate becomes more arid, gypsum and soluble salts in the profile come closer to the surface (to the depth of 2 m). The depth of SOM accumulation is 65-85 cm in the western and 50-75 cm in the eastern parts of the zone. Soil profile is composed of the horizons: $H + H_p + P_{hk} + P_k$. Humic acids dominate in soil humus ($C_{ha} : C_{fa} = 2\text{-}3$, decreasing down the profile to 1.5). The soil reaction is neutral or slightly alkaline (pH_{H2O} 6.5-7.5).

Southern chernozems with low humus content in loess are most widespread in the subzone. They occupy level poorly drained plateaus and their gentle slopes. Their texture is heavy. Soil profile is composed of humus horizon (H) 27-40 cm deep with SOM content of 3-5.5%; dark grey, powdery-cloddy plow layer and grainy subplow layer with abundant earthworm casts; upper transition horizon ($H_{p(i)}$), 10-20 cm thick, dark grey with brownish tint; cloddy; lower transition horizon ($P_{h(i)k}$), 10-20 cm thick; dark brown, nutty with clods, compacted, with streaks and spots of humus-containing material; parent material (P_k) – yellow-brown loess with abundant *bilozirka* within the depth of 65-120 cm and with gypsum crystals and soluble salt efflorescence within the depth of 200-400 cm and deeper.

Southern micelliar-carbonatic chernozems are widespread beyond the Dnister and in the Crimean Steppes. They show effervescence with 10% HCl at a depth of 25-65 cm and carbonate concretions are found at a depth of 65-105 cm. Other forms of carbonate accumulation are pseudomicellium and mold.

Sodicity-affected southern chernozems occur in the areas between chernozemic and arid steppes. They occupy poorly drained areas of Azovo-Prychernomorska (near the Black and the Azov seas) and the Crimean soil provinces. The soil profile differentiation is detected by compactness and nutty-prismatic structure of a transition horizon. The differentiation is proved by the analysis of the soil texture, though it does not effect physico-chemical characteristics. The exchangeable sodium content does not exceed 1-3% of the CEC. Ca/Mg ratio is quite narrow (4.8-.3.3), whereas in nonsodic genuses it is within 4.5-8.2. The soil reaction is close to slightly alkaline (pH $_{H2O}$ 6.5-8.5), soluble salts lie at a depth of 150-250 cm on the left, and 250-350 cm – on the right bank of the Dnipro. The soil profile is composed of the horizons: $H_{(e)}$ + H_{pi} + $P_{hi(k)}$ + $P_{k(s)}$.

Chernosems in non-loess parent materials are found in various zones and subzones. It is difficult to divide them into subtypes within soil subzones and provinces. The factor of parent material "overweighs" the other factors. The soil profile is composed of the horizons: $H_{(k)}$ + $H_{p(k)}$ + P(P_k). There are no visible forms of carbonate accumulation in soil horizons. Only the chernozems formed in certain types of clay can have them. Such soils form in solid clays, ancient sands, eluvium of chalk and chalky marl, eluvium of sandstones, shales, and slates. Very often they are short-profiled or undeveloped.

27.5. Composition and Properties of the Steppe Chernozems

Composition and properties of the ordinary and the southern chernozems are the result of soil humus accumulation, carbonate formation, and the processes of soil salinity and sodicity development taking place in warm climate with inadequate humidity mainly in loess or loess-like parent materials.

Among the ordinary chernozems, heavy loams take up 88.5%, medium loams – 6.5%, clays – 3.9%, and light loams – about 1% of the total area of farmland. Among the southern chernozems, heavy loams and clays occupy 81.6% of the total area of farmland, medium loams – 10.4%, light loams 1.8%, and loamy sands – 1.7%.

Ordinary chernozems do not have any clay fraction redistribution in the profile, whereas the southern chernozems have it, though not too distinctly expressed (Table 86).

The profile differentiation increases to the south and is more expressive in sodicity-affected subtypes. Finely dispersed fraction in the Steppe chernozems is well consolidated with the particles of fine silt forming grainy aggregates not easily destructible in water. Plow layer aggregates are disintegrated into powder, especially in the southern chernozems. As a result, the soils are susceptible to both wind and water erosion.

Table 86. Chemical Composition and Textural Characteristics of Ordinary and Southern Chernozems (M.I.Polupan and V.D.Kysil, 1979)

Characteristics	Ordinary low-humus chernozem in loess			Southern Chernozem in loess		
	Horizons					
	H_a	H	P_{hk}	H_a	$H_{p(i)}$	$P_{hi/k}$
Soil layer, cm	0-20	30-40	90-100	0-10	25-30	40-50
% of particles,						
< 0.01 mm	60.7	58.1	58.2	60.0	59.3	62.2
< 0.001 mm	33.8	33.9	31.3	40.3	42.0	43.1
Oxide content, %						
SiO_2	74.8	75.0	72.4	74.1	73.3	69.0
Fe_2O_3	4.9	4.7	4.7	4.6	4.4	4.6
Al_2O_3	12.8	14.2	12.5	13.5	14.2	13.3
$SiO_2 : R_2O_3$	8.0	7.4	8.0	7.7	7.6	7.2

The mineral composition illustrates the fact that the soil formation occurs without serious changes in the mineral part of the soil. In the southern and eastern directions, as the climate becomes more acid, the depth of SOM accumulation and its content decrease from 125 to 45 cm and from 6.1 to 4.0%, respectively. Soils having lighter texture contain less organic matter (humus). Humic acids dominate in SOM composition, soil humus type is prevailingly humatic (Table 87).

Table 87. SOM Content and Composition in the Steppe Chernozems (M.I.Polupan and V.D.Kysil, 1979)

Soil horizon and depth, cm	Total content of humus, %	Organic C content, %	C_{ha}	C_{fa}	$C_{residual}$	C_{ha} / C_{fa}
			% of organic C content			
Ordinary low-humus chernozem in loess						
$H_{a\ 0\text{-}10}$	5.00	2.90	33.2	13.2	53.6	2.5
$H_{30\text{-}40}$	3.62	2.10	31.7	16.1	52.2	2.0
$H_{p\ 50\text{-}60}$	2.07	1.20	24.8	20.7	54.5	1.2
$P_{hk\ 70\text{-}80}$	1.07	0.60	12.3	33.3	54.4	0.4
Southern Chernozem in loess						
$H_{a\ 0\text{-}10}$	3.10	2.81	30.9	10.7	58.4	2.9
$H_{p(i)\ 25\text{-}50}$	2.60	1.68	30.4	10.1	59.5	3.0
$P_{hi/k\ 40\text{-}50}$	1.50	1.25	22.4	14.4	63.2	1.56
$P_{(h)k55\text{-}60}$	0.90	0.53	19.2	16.4	64.4	1.17

Ordinary and southern chernozems have favorable physico-chemical properties. Their CEC fluctuates within 20-50 meq per 100 g of oven dry soil. Exchangeable Ca and Mg dominate in the soil adsorbing complex (Table 88).

Table 88. Physico-Chemical Characteristics of the Steppe Chernozems (M.I.Polupan and V.D.Kysil, 1979)

Characteristics	Ordinary deep low-humuschernozem in loess				Southern chernozem in loess			
	Horizons							
	H_a	H	H_p	P_{hk}	H_a	$H_{p(i)}$	$P_{hi/k}$	$P_{(h)k}$

Soil layer, cm	0-10	30-40	50-60	70-80	0-10	25-30	40-50	55-60
pH $_{H2O}$	7.2	7.1	7.2	7.5	6.9	7.6	7.6	7.9
Exchangeable cations, meq/100 g:								
Ca^{2+}	31.5	31.4	Not determined	Not deter-mined	25.9	24.0	23.0	Not determined
Mg^{2+}	5.5	7.9			8.3	7.4	7.6	
Na^{+}	0.2	0.2			0.4	0.5	0.6	
K^{+}	0.8	0.5			0.9	0.6	0.5	
CEC, meq /100 g:	37.1	36.4	Not determined		36.4	36.4	34.0	25.0
$CaCO_3$, %	Not determined						2.6	4.6

Exchangeable Na does not exceed 0.5-1.0 meq per 100 g of oven- dry soil. Soil reaction is close to neutrality becoming slightly alkaline down the profile.

CEC is high enough not to limit in any way the productivity of this type of soils soils. Ordinary and southern chernozems are biologically active soils. Bacteria dominate among soil microorganisms. Gypsum and soluble salts in the northern subzone lie at a depth of 3-4 m, in the southern subzone – at the depth of 2-2.5 m, and in the Crimean Steppe – at the depth of 1.5-2 m.

Agrochemical soil characteristics vary depending on soil texture, organic matter content, and the rates of fertilizers applied (Table 89).

Table 89. Soil Fertility Characteristics of Steppe Chernozems (S.P.Latyshev and B.S.Nosko, 1994)

Characteristics	Ordinary chernozem in loess			Southern chernozem in loess		
	Horizons					
	H_a	H	P_{hk}	H_a	$H_{p(i)}$	$P_{hi/k}$
Soil layer, cm	0-20	30-40	90-100	0-10	25-30	40-50

Total (%):						
N	0.27	0.23	0.10	0.16	0.15	0.11
P_2O_5	0.15	0.16	0.13	0.13	0.13	0.12
K_2O	1.9	1.8	2.09	2.6	2.6	2.0
Available forms in Machygin extract, ppm:			Not determined			Not determined
P_2O_5	37.8	24.3		32.6	21.3	
K_2O	350	278		386	303	

By the indices of soil fertility, both chernozems are well supplied with available P_2O_5 and K_2O. Water related and physical properties of the steppe chernozems are, generally, not bad (Table 90). But down the profile, the bulk density of the soil and its total porosity may be slightly less than favorable, so that carbonatic and illuviated horizons may have poor penetrability for the roots of certain plants.

Table. 90. Water-Related and Physical Properties of the Steppe Chernozems (M.I.Polupan, V.D.Kysil and P.O. Gavrik, 1979)

Characteristics	Ordinary deep low-humus chernozem in loess				Southern chernozem in loess			
	Horizons							
	H_a	H	H_p	P_{hk}	H_a	$H_{p(i)}$	$P_{hi/k}$	$P_{(h)k}$
Soil layer, cm	0-10	30-40	50-60	70-80	0-10	25-35	40-50	55-60
Bulk density, g/cm^3	1.17	1.19	1.22	1.36	1.00	1.26	1.32	1.42
Total Porosity, % of volume	56.2	54.6	53.8	48.9	61.6	51.6	49.9	46.3
PWP, % m	12.1	12.4	12.3	11.7	11.0	11.1	12.9	12.3
FC, % m	29.8	24.4	23.4	22.6	31.9	27.9	23.9	23.3
Active moisture range, mm in 10 cm layer	20.7	14.3	13.6	14.8	20.9	21.1	14.6	15.6

The amount of available water at field capacity (FC) in the upper 1 meter layer of medium loams is within 160-180 mm, but this is only a potential soil characteristic, whereas in actual fact the amount of available water is determined by the climate's aridity. Having evaluated potential productivity of ordinary chernozems according to a 100-point scale, A.I. Siry found it to be within 55-86 points. The greatest potential productivity is characteristic of deep moderately humusified ordinary chernozems, heavy loams and light clays by texture. Southern chernozems proved potentially less productive (49-56 points).

27.6. Agricultural Use of the Steppe Chernozems and Practices of Conservation and Improvement

Soil moisture is the main limiting factor of soil productivity in the Steppe zone. All agronomic practices aimed at water harvesting, conservation and economical use and the practices of irrigation, wherever it is feasible, are of paramount importance. Summer fallow may be effective on the ordinary and the southern chernozems. Such practice allows to accumulate sufficient amounts of water for high and stable yields of crops compared with nonfallow preceding fields.

Periodic droughts in the Steppe zone come 2 to 6 times in a decade. To combat droughts, the surface runoff must be turned into a subsoil one. The contour organization of farmland areas is a necessity in the zone. Much is being said about it but the transition to the contour farming has not been made to this day. The fields remain rectangular. Contour direction of tillage practices, shelterbelts, terraces, diversions, etc. will help to conserve water and soil, curb erosion and sedimentation. Row crops should not be allowed on the slopes over 3°, and land areas on the slopes over 5° should be turned into grasslands or used for the purposes of afforestation.

The non-plow tillage is less destructive to the soil structure and agrophysical conditions (tilth), than moldboard plowing. The non-plow tillage operations with V-blade cultivators and disks are more fuel and labor-saving. The conservation tillage should allow to leave over 30% of soil surface covered with the residue of a previous crop during planting of some definite crop.

The restricted tillage up to the zero one, still seems rather hazardous to many Ukrainian farmers. More research is needed in this sphere. The non-plow tillage operations leaving residues on the soil surface are an effective prevention of dust storms and less severe forms of wind erosion.

The annual loss of SOM from a hectare of plowland in the Steppe zone of Ukraine is now 0.4-0.5 mt. To avoid this loss it is necessary to apply additional farm manure in rates of 6-8 mt per hectare, of a crop rotation per year, whereas the total rate of FM per hectare of a crop rotation per year should be 8-10 mt. Soil humus allows to protect soil tilth from rapid ruination.

The irrigation in the Steppe zone is a necessity, but it must be strictly regulated and accompanied by attentive and timely soil monitoring to prevent soil degradations related to irrigation, like the growth of salinity and sodicity, irrigation erosion, as well as soil pollution, waterlogging and swamping. Irrigation water quality criteria should be derived from the characteristics of both, water and soil, forming a single system on irrigated lands. Besides, winter wheat, corn, barley, vineyards, and orchards plants, the steppe chernozems (ordinary) are used for the production of sugar beet.

Questions & Assignments

1. Speak on the geographical position and the area of the Steppe zone.
2. Which specific features characterize the soil formation factors in the Steppe zone?
3. How does climate effect the natural vegetation and the soil in the Steppe zone?
4. Speak on the parent materials in the Steppe zone.
5. Name the soil formation processes in the Steppe zone.
6. Speak on the Steppe zone chernozems classification.
7. Name the essential differences between ordinary and southern chernozems.
8. Describe micelliar-carbonatic genuses of ordinary and southern chernozems.
9. Describe chernozems in non-loess parent materials.
10. Describe physico-chemical characteristics of ordinary chernozems.
11. Speak on soil degradations in the Steppe zone.
12. Propose soil and water conservation practices for the Donetsky ridge area.
13. Which factors limit the potential productivity of the Southern chernozems?
14. What practices should be used to prevent soil organic matter losses in the Steppe zone of Ukraine?
15. Which crops are most suitable for production in the northern subzone of the Steppe zone of Ukraine and why?

Chapter 28

Soils of the Arid Steppe Zone of Ukraine

The Arid Steppe zone occupies a band-shaped area, the width of which varies from 5 to 150 kilometers. It includes the central (lowest) part of the Prychornomorska lowland, “embracing” the Black sea from the north, the Sivash, and the northern (lowland) area of the plain land in Crimea. The middle part of Odesa, southern parts of Mykolaiv and Kherson regions, and the north of the Autonomic Republic of Crimea make up the zone, the total area of which is 4,711 thousand hectares. A boundary area between the Steppe and the Arid Steppe zones, is not wider than 20-25 kilometers. Within this boundary band the sodicity-affected southern chernozems, formed on the plateaus, change into dark chestnut sodicity-affected soils. The eroded southern chernozems still dominate on the slopes. Chestnut soils occupy the lowest part of the Prychernomorska lowland bordering with the Sivash coasts, as well as the coasts of the Black and the Azov seas. On the Crimean peninsula, the Arid Steppe zone is followed by the Steppe zone in the south. Some soil geographers explain this inversion of soil zones by the effects of the Crimean mountain system with its vertical zonality of climate.

28.1. Factors of Soil Formation

Climate. By climatic conditions, the Arid Steppe zone with dark chestnut and chestnut soils, differs from the other zones of the Ukrainian Steppe by the greatest resources of heat and aridity. The sum of temperatures over 10°C is in the range of 3,400-3,600°C. The mean monthly temperature of July is 26-27°C and that of January – minus 2.2-3.0°C. The mean annual sum of precipitation is in the range of 300-350 mm, but in droughty years it does not exceed 250 mm. The rains often have the character of heavy showers. Humidity coefficient fluctuates within 0.3-0.6. Dry eastern winds in summer strengthen the droughts and may cause dust storms. An average annual rainless period lasts for 25-45 days, once in five years it lasts for 40-60 days, and once in a decade – for 50-70 days.

High air temperature and low air humidity with prolonged rainless periods initiate droughts and severe dry winds damaging the crops. Summer droughts occur every year during the growth of late-maturing crops. Spring droughts have a local character. But the amount of available water in soil in spring is usually low and the yields of crops depend on the rains in summer.

The Arid Steppe is the zone of hazardous farming. The relief of the Arid Steppe is represented by a plain land sloping to the south and further to the north, to the Sivash shores, and the to coasts of the Black and the Azov seas. Only in the area between the Dnister and the Dnipro there are long ravines, 20-30 m in depth, with gentle slopes. Absolute altitudes are within 40-50 meters. The surface of loess terraces is flat but covered with numerous pods and depressions. Sand terraces have dunes and ridges of sandy deposits. The Crimean part of the zone is most level and poorly drained. Flat plain land here gradually rises to the south from 5-10 to 50 meters above sea level. Ground water table is within 1.5-2.0 meters from the surface. Ground water is characterized by considerable salinity. Micro and nano-land forms in the zone influence the development of salinity and sodicity in soils.

Vegetation. Natural vegetation in the Arid Steppe zone is represented by a poorly developed association of grasses which include feather-grass and needle-grass, as well as fescue and wormwood. Soil surface is covered with such grass by only 50-70% of its area, as the greater the aridity is the thinner is the cover. The role of ephemeroid species, like tulips and peonies is considerable.

Near the Sivash and on the coasts of the Black and the Azov seas, the natural grass associations contain numerous halophytic species like salsola and suede grass, as well as some varieties of wormwood. The soil surface may be covered with lichens and algae, mainly green-blue and diatomic.

Plant biomass in the Arid Steppe zone is in the range of 15-20 mt/ha, over 90% of which is the biomass of roots. The annual increment of the above-ground biomass amounts to 2.5-3.0 mt/ha, and that of the roots – 10-11 mt/ha. Every year, up to 0.6 mt/ha of ash elements and 0.1-0.15 mt/ha of nitrogen become involved in the biologic cycle. The return of the nutrient elements to the soil is approximately equal to their consumption. The biologic activity in chestnut soils is weaker than in chernozems.

Parent materials in the Arid Steppe are represented by heavy-textured loess-like loams and loess. The depth of loess deposits on the plateaus may be within 20-25 m. Alluvial deposits and clays are parent materials only on small areas. Still smaller area is occupied by eluvium-deluvium of sandstones, shales, limestones, and marl. They are found on the right-bank slopes of river valleys and ravines. Floodplain soils form in alluvial deposits of varying chemical and mechanical composition. The salt-bearing horizon of sulfate, chloride, sodium, and gypsum accumulation lies within 1.7-2.2 m from the soil surface.

Pods (saucer-shaped depressions) are formed by gleyed loess deposits. Their texture is medium and heavy loam, as a result of gleying. In the northern part of the zone, the loess deposits in the pods are leached from soluble salts, but on the Sivash shores and the Black Sea coasts loess in pods is saline. The same pertains to ground water, which is nonsaline in the north and very saline in the south. On the areas among the pods, ground waters lie in the lower portion of loess deposits and are saline everywhere.

28.2. Soil Genesis in the Arid Steppe Zone

Chestnut soils were formed in arid climate, on the level topography, under xerophyte grass associations, mainly in carbonate-containing parent materials, containing a certain amounts of soluble salts. Soil formation includes three essential processes: 1) humus accumulation, 2) carbonate formation, and 3) development of sodicity. The leading of these processes is the accumulation of humus and biogenic elements in the horizons of SOM accumulation, which in its essence, is a sod-forming process. But sod-forming process in the Arid Steppe is different from the same process in the Steppe and Forest-Steppe zones. The first specific feature is poor vegetation with halophytic species. Ephemeroidal plants have weakly developed roots. The depth of humus accumulation is smaller. SOM content and amount in a layer of humus accumulation are considerably lower than in chernozems. Humus is more fulvic in composition. Droughty climate impedes soil humus accumulation. The annual contribution of plant residues to the soil is 4-8 mt/ha. This means the involvement of 250-300 kg of nutrient elements in the biologic cycle.

The process of carbonate formation (carbonatization) is a common feature of chernozems and chestnut soils. The chemical equation describing calcium carbonate precipitation may be written in the following way:

$$Ca^{2+} + H_2O + CO_2 \rightarrow CaCO_3 + 2H^+$$

The dissolution of $CaCO_3$ may follow the equation:

$$CaCO_3 + CO_2 + H_2O \rightarrow Ca(HCO_3)_2$$

Carbonates in dark chestnut and chestnut soils, like those in the ordinary and southern chernozems, form soft concretions (nodules) which are called *bilozirka* in Ukrainian. In chestnut soils, alongside the carbonates are sometimes found the crystals of gypsum and soluble salts. There is not enough moisture to leach them down. The decomposition of natural vegetation residues releases a considerable amount of sodium, which causes the development of sodicity.

The development of sodicity is one of specific features of soil genesis in the Arid Steppe. The saturation of soil adsorbing complex (SAC) with exchangeable sodium causes the transformation of secondary minerals and dispersion of humus and other colloids, some part of which move downward in the profile causing its differentiation into eluvial and illuvial horizons. In the last decade or so, Ukrainian soil scientists speak about physical sodicity versus the sodicity caused by more than 5% of exchangeable sodium of CEC. Physical sodicity demonstrates itself in the differentiation of soil profile into eluvial (H_e) and transition illuvial (H_{pi}) horizons. The first of them is platy, the second – nutty prismatic. In the majority of chestnut soils, their sodicity is not caused by the sufficient for this purpose content of exchangeable sodium.

The soil cover in the Arid Steppe zone is very complex, characterized by the presence in the soil complexes of mapping units with different extent of sodicity and salinity accompanied by different extent of soil hydromorphicity. Many researches explain this phenomenon by the complexity of micro land forms and hydrologic conditions.

28.3. Essential Soil Types, Their Diagnostics and Agronomic Properties

Dark chestnut sodicity-affected soils occupy an area of 1,207 thousand hectares of which 1,089 thousand hectares are a plowland. Chestnut soils occupy an area of 219 thousand hectares of which 151 thousand hectares are a plowland. There are also meadow-chestnut sodicity-affected soils, the total area of which reaches 62 thousand hectares including 49 thousand hectares of a plowland. Characteristic depressions (pods) occupy 10.8 thousand hectares. The soils in pods are mainly meadow-chestnut gleyed from the surface or just gleyed, among which there are found solodized soils (gleysolods). Many soils are eroded. Solonchaks, solonetzes and solods are found too.

28.3.1. Chestnut Soils Classification

Chestnut soils and their classification are named in FAO nomenclature as *Castanozems haplic***.** According to soil classification (Table 91), by the combined depth of H_e and H_{pi} horizons, chestnut soils are divided into four subtypes: dark chestnut sodicity-affected, chestnut sodicity-affected, dark chestnut secondarily sodicity-affected from the surface, and chestnut secondarily sodicity-affected from the surface.

Table 91. Classification of Chestnut Soils in Ukraine (Ukrainian Research Institute of soil science and Agrochemistry, 1988)

Subtype	Genus	Species
Dark chestnut Dark chestnut sodicity affected Chestnut sodicity-affected Dark chestnut and chestnut secondarily sodicity affected from the surface	Modal, carbonatic, eroded, washed-on, blown-on, deeply plowed, secondarily saline, secondarily meadowish, secondarily meadowish saline, mocharic	By the extent of sodicity, erosion, salinity, depth of the layer of SOM accumulation, humus content, depth of carbonates, depth of watertable, upper salinity affected horizon, extent of washing-on or blowing-on, etc.

Secondarily sodicity-affected from the surface soils have, as a result of irrigation, a structureless and slaty plowlayer, which, when wet, turns into a sticky mass. When dry, it is also very compacted and hard.

The most widespread genuses of the chestnut soils are:

1) **modal (ordinary):** as the name implies, with the most typical characteristics;
2) **sodic or sodicity-affected**: within the layer of SOM accumulation they have a sodic (solonetzic) horizon with an exchangeable sodium content over 3% of the CEC;

3) **carbonatic (calcareous):** show effervescence with dilute HCl from the very surface and along the entire profile;

4) **eroded:** differ from the modal genuses by a partial or complete loss of the upper horizons, as a result of water or wind erosion;

5) **washed on or blown-on:** that is, buried by a layer of erosion sediment. The surface horizon is 30 or more centimeters deeper and often layered (platy);

6) **with secondary salinity**: occur in irrigated areas. Soluble salts are concentrating in a soil profile and may form salt crusts on the surface;

7) **secondarily meadowed (*secondary,* resulting from irrigation):** occur in irrigated or waterlogged areas; the soils with ground water within 3-5 m of the surface are **meadowish,** whereas those with ground water closer than 3 m to the surface are **meadow**.

The species of chestnut soils are identified by:

– the extent of soil sodicity (% of exchangeable Na^+ of the CEC):

1) slightly sodicity-affected: 1-3%;
2) moderately sodicity-affected: 3-6%;
3) strongly sodicity-affected: 6-10%,
4) sodic (solonetz soils): over 10%;

– the extent of erodedness (xeromorphicity):

1) **slightly eroded:** up to ½ of $H_{(e)}$ horizon or up to 20% of SOM lost;

2) **moderately eroded:** from ½ to the whole of $H_{(e)}$ horizon and from 20 to 40% of the initial amount of SOM lost,

3) **severely eroded:** lost $H_{(e)}$ and the next horizon, partly or completely; 40-60% of the initial amount of SOM lost from the soil;

– by the extent of salinity the soils may be slightly, moderately, and very saline;

– by the depth of the layer of SOM accumulation:

1) deep: $H_e + H_{pi} > 50$ cm;
2) averagely deep: $H_e + H_{pi} = 30-50$ cm,
3) not deep: $H_e + H_{pi} < 30$ cm;

– by SOM content in the surface layer:

1) with slight SOM content: < 3% and
2) with low SOM content: > 3%;

– by the depth of the upper saline horizon:

1) solonchaky: 0-30 cm;

2) highly solonchaky: 30-50 cm;
3) solonchakish: 50-100 cm;
4) deeply solonchakish: 100-150 cm,
5) nonsolonchaky: over 150 cm deep;

– by the depth of the effervescence line:

1) highly carbonatic: line in the plow layer,
2) deeply carbonatic: line in the parent material;

– by the depth of ground water table:

1) gleyish: 1.5-2.0 m;
2) gleyed: 1.0-1.5 m;
3) very gleyed: 0.5-1.0 m,
4) gleyed from the surface;

– by the depth of washing-on and by the depth of blowing-on:

1) slightly (washed-on or blown-on) up to 25 cm;
2) moderately (washed-on or blown-on) 25-50 cm,
3) severely (deeply) (washed-on or blown-on): > 50 cm.

There are also lithologic series, variants (virgin, reclaimed, cultured, irrigated, and drained), and varieties by texture.

28.3.2. Chestnut Soils Morphological Features

Dark chestnut sodicity affected soils occur in the Arid Steppe zone mainly on poorly drained plain lands of the watershed between the Dnipro and the Molochna and in the northern part of the Crimean Steppe. Some small areas of such soils occur on the lowland seaside plateaus in the right bank part of the zone. The soils were formed under the associations of needle grass, feathergrass, cereal grasses and warmwood. Their profile differentiation is identified both morphologically and by the particle-size analysis in the laboratory. The layer of soil humus accumulation is 50-60 cm thick. Gypsum and soluble salts lie below the depth of 150-250 cm. Right-bank soils have them in deeper layers.

Irrigated variants of dark chestnut soils possess some specific cloddiness and blockiness in the plow layer and a considerable bulk density. After rains, they become puddled and form quite a firm crust on the surface. On considerable areas, the dark chestnut soils are deeply plowed and contain 10-30% less SOM than their virgin counterparts.

A typical soil profile of the dark chestnut soil has the following description:

H_e: slightly eluviated horizon of SOM accumulation; 25-40 cm thick; dark grey with chestnut shade; powdery with clods; blocky when dry; faces of soil aggregates “dusted” with amorphous SiO_2; transition may be clear (by the plow toe) or gradual;

Hpi: upper transitive illuviated horizon; 10-30 cm thick; dark chestnut; contains clods, grains, and nuts, or clods and prisms with nuts; compacted; amorphous silica on the faces of aggregates in the upper portion of the horizon; slight colloidal “luster” on the same faces in the lower portion of the horizon; earthworm activity traces; transition gradual;

Phik: lower transition horizon; 10-30 cm thick; dark brown, may be with dark grey spots of humus-containing material; nutty with prisms and / or clods; effervesces with 10% HCl; colloidal films on the aggregate faces; transition either gradual or wavy with tongues;

Pk: parent material; loess with carbonate concretions (bilozirka) within the depth of 50-120 cm and crytalls of $CaSO_4 \cdot 2H_2O$ within the depth of 150-250 cm.

Moderately and severely sodicity-affected soils occupy the elevated elements of topography. Their profiles are more distinctly differentiated into eluvial and illuvial horizons. Thus, a severely sodicity-affected soil may have the horizons: HE + HI + Phi + Pks.

On well-drained areas in the western part of the zone, the nonsodic dark chestnut soils may be found. Their profile has the following horizons:

H: horizon of humus accumulation; 20-25 cm deep; darkish grey with a slight brownish hue; loose, friable; powdery with grains; much transformed by the earthworm activity; transition gradual;

HP: transition horizon 35-40 cm thick; brownish grey; coarse-grained; loose; rare crotovinas; rapid transition to parent material;

$P_{/k}$: parent material; loess; medium loam; carbonate nodules (bilozirka) within the depth of 75-100 cm. Gypsum and soluble salts are not detected within the soil profile.

Such soils may contain 2-2.5% of SOM and have CEC within 10-15 meq/100 g. In humid years or with irrigation the soils may prove quite productive.

Chestnut slightly sodicity-affected soils: found in mosaics with sodic and dark chestnut sodicity-affected soils. They effervesce from the depth of 40-55 cm. The profile description is as follows:

H_e: slightly eluviated horizon of SOM accumulation; 25-30 cm deep; dark brown; when dry – brownish grey; powdery with clods; the entire horizon is involved in a plow layer; transition distinct along the plow toe.

H_{pi}: upper transition horizon; 10-15 cm thick, dark cinnamon; when dry – cinnamon brown; nutty with clods and grains; loose; transition gradual;

$P_{hi(k)}$: lower transition horizon; 10-15 cm thick; dark cinnamon (chocolate); nutty with grains and clods; slightly compacted; carbonatic more often than not; transition gradual;

P_k: parent material; loess slightly “soiled” with humus-containing material in the upper portion; carbonate nodules (bilozirka) within the depth of 70-100 cm; gypsum and soluble salts embed below the depth of 100-150 cm.

The profile of moderately sodicity-affected soil is like this: $H_e + H_{pi} + P_{hi/k} + P_k$, whereas that of severely affected is the following: HE + HI + $P_hi_{/k} + P_k$.

28.3.3. Chestnut Soils Composition and Properties

By texture, heavy loams and light clays are dominant among sodicity-affected chestnut soils. The surface $H_{(e)}$ horizons contain less physical clay than $H_{p(i)}$ and P_{hik} horizons (Table 92).

Table 92. Chemical Composition and Textural Characteristics of Chestnut Soils (M.I. Polupan, 1979)

Characteristics	Dark Chestnut Sodicity-Affected Light Clay Soil in Loess				Chestnut Sodicity-Affected Light Clay Soil in Loess			
	Horizons							
	$H_{(e)a}$	$H_{p(i)}$	$P_{hi/k}$	$P_{k(h)}$	H_{ed}	HP_{ik}	P_{hik}	$P_{k(h)}$
Soil layer, cm	0-10	35-45	50-60	70-80	0-6	17-43	33-43	85-95
Content of particles, %								
< 0.01 mm	60.9	63.5	61.5	63.6	61.0	63.1	62.5	58.1
< 0.001 mm	38.4	39.8	40.0	42.4	40.2	42.2	44.9	41.6

Oxide content, %								
SiO_2	71.4	98.7	68.5	68.4	65.3	59.7	56.3	68.9
Fe_2O_3	5.4	6.8	6.2	5.0	3.9	5.8	5.2	5.1
Al_2O_3	9.0	11.1	10.3	12.4	7.5	9.6	8.7	8.4
$SiO_2:R_2O_3$	5.0	3.8	4.2	3.9	5.7	3.9	3.8	5.1

The fraction < 0.001 mm is dominated by the secondary minerals like montmorillonite and hydromicas. The total content of SiO_2 in the soil profile changes little. R_2O_3 content is determined by the extent of soil sodicity and increases in illuvial horizons. SiO_2/R_2O_3 ratio is the highest in the eluvial portion of soil profile (5.0-5.7) and reduces to 3.8-4.2 in the illuvial (sodic) horizons.

Humus in chestnut soils is firmly associated with the mineral part of the soil. In dark chestnut soil, SOM (humus) content may reach 2.5-3.5%, whereas in chestnut soils it is within 1.3-3.0%. Soil reaction in dark chestnut soils is close to neutral, whereas in chestnut soils it is slightly to moderately alkaline, which is caused by the presence of exchangeable sodium and soluble sodium salts. With depth, the carbonate content and soil alkalinity increase (Table 93).

Table 93. Physico-Chemical Characteristics of Chestnut Soils (M.I.Polupan, 1979)

Characteristics	Dark Chestnut Sodicity-Affected Light Clay Soil in Loess				Chestnut Sodicity-Affected Light Clay Soil in Loess			
	Horizons							
	$H_{(e)a}$	$H_{p(i)}$	$P_{hi/k}$	$P_{k(h)}$	H_{ed}	HP_{ik}	P_{hik}	$P_{k(h)}$
Soil layer, cm	0-10	35-45	50-60	70-80	0-6	17-43	33-43	55-65
pH H_2O	6.5	6.8	7.0	7.2	7.7	7.8	8.0	7.9

Exchangeable cations, meq per 100 g of oven dry soil:								
Ca^{2+}	23.7	24.0	22.4	Not deter-mined	16.7	20.9	Not deter-mined	Not deter-mined
Mg^{2+}	6.9	7.4	8.4		8.0	10.7		
Na^{+}	0.5	0.6	0.7		1.0	1.3		
K^{+}	1.1	0.9	0.7		1.0	0.8		
CEC, meq per 100 g of oven dry soil:	34.0	33.4	32.4		26.5	29.6	30.3	
Exchangeable Na^{+} content, % of the CEC:	3.2	2.7	2.2	Not deter-mined	3.8	2.7	Not deter-mined	Not deter-mined

The soil adsorbing complex of chestnut soils is saturated with calcium and magnesium cations. Ca/Mg ratio is low (3.7-2.1), which may cause magnesium sodicity. Dark chestnut soils of the right-bank part of the Arid Steppe zone are nonsaline. Total dissolved solids do not exceed 0.07-0.14% (Table 94).

Table 94. Soil Extract Analysis Results for Chestnut Soils (Data of M.I.Polupan in meq per 100 g of dry soil)

Characteristics	Dark Chestnut Sodicity-Affected Light Clay Soil in Loess				Chestnut Sodicity-Affected Light Clay Soil in Loess			
	Horizons							
	$H_{(e)a}$	$H_{p(i)}$	$P_{hi/k}$	$P_{k(h)}$	H_{ed}	HP_{ik}	P_{hik}	$P_{k(h)}$
Soil layer, cm	0-10	35-45	50-60	70-80	0-6	17-43	33-43	85-95

TDS, %	0.07	0.08	0.1	0.14	0.12	0.21	0.26	0.41
Anions: CO_3^{2-}	Not detected			0.17	Not detected			0.12
HCO_3^-	0.55	0.85	1.09	1.44	0.52	0.79	1.68	0.80
Cl^-	0.1	0.06	0.07	0.09	0.04	0.03	0.68	3.42
SO_4^{2-}	0.2	0.05	0.09	0.14	1.07	2.09	2.37	2.09
Cations:								
Ca^{2+}	0.69	0.56	0.32	0.37	0.47	0.47	0.56	0.32
Mg^{2+}	0.11	0.07	0.31	0.04	1.03	1.96	2.24	0.3
Na^+	0.09	0.28	0.62	1.34	0.18	0.26	1.32	5.3
K^+	0.06	0.001	0.01	0.01	0.05	0.005	0.005	0.008

With a relatively low SOM content, the total nitrogen content in a plow layer of dark chestnut soils is about 0.15%, whereas in chestnut soils it is about 0.13% (Table 95). Chestnut soils contain more total P_2O_5 (0.10-0.13%) compared with dark chestnut soils (0.09-0.11%). Both dark chestnut and chestnut soils contain much total K_2O (2.1-2.5%).

Table 95. Soil Fertility Characteristics of Chestnut Soils (B.S.Nosko and S.P.Latyshev, 1994)

Characteristics	Dark Chestnut Sodicity-Affected Light Clay Soil in Loess			Chestnut Sodicity-Affected Light Clay Soil in Loess		
	Horizons					
	$H_{(e)}$	$H_{p(i)}$	$P_{hi/k}$	H_e	HP_{ik}	P_{hik}
Soil layer, cm	0-10	35-45	50-60	0-6	17-33	33-43

Total Content, %						
of N	0.15	0.12	0.09	0.13	0.1	0.07
P_2O_5	0.11	0.09	0.07	0.13	0.1	0.08
K_2O	2.4	2.1	2.5	2.1	2.1	2.3
Content of available forms in Machigin extract, ppm						
P_2O_5	21.6	18.3	Not determined	8.9	20.6	Not determined
K_2O	267	225		30	298	

Both dark chestnut and chestnut soils are moderate to well supplied with available P_2O_5 (18-29 ppm). Exchangeable K_2O content is optimal in both soils (225-305 ppm). Microelements are contained in above-average levels: Co (17-18 ppm), Mn (560-1200 ppm), Zn (110-140 ppm), B (30-40 ppm), and Cu (30-60 ppm).

Dark chestnut and chestnut soils have satisfactory physical and soil water related properties, close to those of southern chernozems. The bulk density of plow layers is within 1.15-1.17 g/cm^3, which is favorable for most crops. In transitive illuviated and, particularly, carbonatic horizons the bulk density of the soil grows to 1.40-1.48 g/cm^3 (Table 96). Soils with secondary sodicity are overcompacted in sodic horizons.

Table 96. Physical and Soil Water Related Properties of Chestnut Soils (M.I.Polupan, et al., 1994)

Characteristics	Dark Chestnut Sodicity-Affected Light Clay Soil in Loess			Chestnut Sodicity-Affected Light Clay Soil in Loess		
	Horizons					
	$H_{(e)}$	$H_{p(i)}$	$P_{hi/k}$	H_e	HP_{ik}	P_{hik}
Soil layer, cm	0-10	35-45	50-60	0-6	17-33	33-43
Particle density, g/cm^3	2.64	2.65	2.68	2.53	2.67	2.67

Bulk density, g/cm^3	1.17	1.32	1.48	1.15	1.24	1.40
Total Porosity, % of volume	55.7	50.2	44.8	56.3	53.6	47.6
PWP, % of dry weight	4.2	15.0	15.3	3.5	17.1	15.9
FC, % of dry weight	26.0	22.0	21.0	29.0	26.1	24.2
Available water at FC, % of dry weight	11.8	7.0	5.7	15.5	9.0	8.3

The soil aggregation in chestnut soils is satisfactory but not so good as in southern chernozems. The soil structure deteriorates with irrigation. Plow layers become blocky. Illuvial horizons become impenetrable to water and eluvial – slaty, with low aeration porosity. Infiltration coefficients do not exceed 0.2-0.3 mm per minute. The autumn rains do not soak the soils deeper than 70-100 cm. Snowmelt water and rains in spring can make soil wet to the depth of 1.5-2 m.

Chestnut soils generally are susceptible to wind erosion. Most rich in SOM are dark chestnut soils (230 mt/ha) in the northern part of the zone. Such soils are according to least sodicity-affected. Their potential productivity by 100-point scale may reach 37 points. Chestnut sodic soils may be evaluated as having 20 points of potential productivity (A.I.Siry, 1998).

28.3.4. Agricultural Use and Practice of Chestnut Soils Improvement

The main reasons of the low productivity of dark chestnut and chestnut soils are: storage of water, droughtiness of climate, physical and chemical sodicity, salinity, alkaline soil reaction, erodedness, low humus content, not too good physical properties, and in some regions – soil skeletality (stonyness). The depth of plow layer and the presence of plow toe limit the productivity of chestnut soils. There is a need to periodically involve the transition horizon into a plow layer. The negative consequences may be the growth of sodicity and salinity in the layer. Numerous agricultural crops are cultivated in the Arid Steppe zone: winter and spring wheat, corn for grain, rice, barley, millet, soybeans, sunflowers, watermelons, melons, fruits, and grapes. Of great importance here are the practices of soil and water conservation, as well as irrigation. The operation of planting should be performed a very limited period of time.

Dark chestnut and chestnut soils with low sodicity and not too well expressed illuvial horizons rich in colloids, subsoiling to the depth of 28-30 cm improves water regime. Chisel plows are the best for a primary soil tillage, as they can loosen hard soil layers. At the same time, this kind of tillage implements helps to protect the soil from wind erosion. The soil water is better conserved.

In soils with the great extent of sodicity, containing over 6-10% of Na^+ of the CEC, a thick illuvial horizon is formed at the depth of 20-40 cm. It is impenetrable to water and air and impedes the growth of roots. To improve the productivity of such soils, the illuvial (sodic) horizon must be broken by deep plowing (to the depth of 60-70 cm). Calcium carbonate and partly gypsum from the lower horizons improve the physico-chemical properties of the entire layer of root penetration. But the lower soil layers turned to the surface contain less SOM and are characterized by a lower fertility in general. Deep plowing must be combined with the application of gypsum and organic manures. Perennial grasses might improve soil conditions.

In the areas with high wind erosion hazard, on the windward slopes, soil-conservation tillage and special crop rotations are a must. Non-plow tillage to the depth of 20-22 cm should be primary for winter wheat. Row crops demand a deeper nonplow tillage (to the depth of 25-28 cm), whereas spring crops after intertilled crops in a rotation may need a shallow (10-12 cm) primary tillage. The barriers of tall-grown crops (corn, sunflowers) are a good addition to windbreaks in the soil protection from blow-out. The rows of plants should be arranged across the direction of the wind. The contour strip cropping is effective in the water erosion control. Non-plow implements, leaving some stubble on the soil surface, help to protect the soil against both wind and water erosion. Such practice also allows to trap snow in winter and provides the soil with more water in spring. To combat the effects of soil sodicity, a regular application of gypsum $(CaSO_4 \cdot 2H_2O)$ should be within the rates of 2-3 mt/ha.

The most effective fertilizers in the Arid Steppe zone are nitrogen and phosphorus ones. Potassium fertilizers are least effective for most crops. But with irrigation, even potassium may appear deficient. It is important to use correct rations between the elements of nutrition (not more than $N_{20-60}P_{10-40}K_{20}$) and apply them under most sensitive and priority crops like winter wheat, especially after non-fallow preceding crops, spring and winter barley, corn, sunflower, and rice. The other crops “utilize” the aftereffects of fertilizers applied for the previous crops.

Farm manure should be applied at rates within 6-8 mt per hectare of a crop rotation per year. This ensures a non-deficit balance of SOM.

Irrigation should compensate the consumption of water by the plants and the loss of water through evaporation. With the expenditures of water at rates of 10 mm per 0.1 mt of grain per hectare, the yields of small grain crops without irrigation even in years without severe droughts can be only within 2-3 mt/ha. Irrigation water quality must be evaluated not only by the percentage of TDS and alkalinity characteristics. The ratios between Ca, Mg, and Na should be taken into account as well as irrigated soil characteristics. Irrigation techniques should be more up-to-date.

Chestnut and very sodicity-affected soils in mosaics with sodic and very saline areas should be used for pastures with the regulated (controlled) grazing.

Questions & Assignments

1. Show the Arid Steppe zone on a soil map of Ukraine and tell about its natural features.
2. Describe the factors of soil formation in a soil zone.
3. Compare natural and cultivated vegetation in the Arid Steppe zone. How does vegetation effect the nature of soils?
4. Speak on the parent materials in the Arid Steppe zone and their effects on the soils.
5. Which processes of soil formation are essential in the Arid Steppe zone?
6. What is the difference between physical and chemical sodicity?
7. Speak on chestnut soils classification.
8. Describe morphologic features of sodicity-affected chestnut soils.
9. Describe physico-chemical and general physical properties of sodicity-affected chestnut soils.
10. Name the factors limiting potential productivity of chestnut soils. Explain how they limit it.
11. Name the practices of soil conservation in the Arid Steppe zone.
12. Name the practices of soil improvement in the Arid Steppe zone.

Chapter 29

Salt-Affected Soils in Ukraine

Salt-affected soils include a large group of soils of different genesis and characterized by different properties. But all of them have some relation to soluble salts in their profile, which either are contained in the soils in quantities deteriorating their properties and productivity or influenced in some way the origin of a given soil.

As long ago as in 1964, the International Society of Soil Science set up a Subcomission on Salt-Affected Soils with a view to identifying and locating the salt affected soils of the world. The first draft of the Map of European Salt Affected Soils was presented during the Symposium of the Subcomission on Salt-Affected Soils of the International Soil Science Society in Yerevan, Armenia, in May, 1969. The second, revised map, was discussed by the working group in Smolenice, then Czechoslovakia, in February, 1970, and was accepted in its final form during the next meeting of the Subcomission in Sevilla, Spain, in May, 1971. It is generally accepted, that watersoluble salts, particularly the sodium salts, are responsible for the low productivity of salt-affected soils.

Salt-affected soils are common in arid and semiarid regions, where annual precipitation is insufficient to meet the evapotranspiration needs of plants. As a result, salts are not leached from the soil. Instead, they accumulate in amounts or types detrimental to plant growth. Salt problems are not restricted to arid or semiarid regions. They can develop even in subhumid and humid regions under appropriate conditions. Salt-affected soils often occur within irrigated lands, the area of which in Ukraine reaches 2.6 mln hectares.

29.1. Essential Notions and Definitions

Solonchaks (the Ukrainian term for very saline soils) contain in the surface (0-30 cm) layer of soil over 1% of chlorides, or over 2% of sulfates. Soluble salts possess different toxicity. Sodium carbonate (Na_2CO_3) is the most toxic for plants. When its content exceeds 0.1% the growth of plants becomes impaired, whereas with 0.6% and more, the soil becomes barren. Solonchak profiles do not have their special horizons. They preserve relic horizons of a zonal soil subjected to such a severe extent of salinity.

Solonetz (sodic) soils contain sufficiently large amount of exchangeable sodium (and sometimes magnesium) in illuvial (sodic) horizon. Sodic soils have a sharp differentiation of a soil profile into eluvial and illuvial horizons. They contain soluble salts not in the surface horizon but leached to a certain depth.

Solod (Solot in the Canadian System of Soil Classification) soils are hydromorphic or semihydromorphic soil species with distinct eluvio-illuvial differentiation of a profile. Their eluvial (solodized) horizon is whitish in color due to the abundance of amorphous silica. According to K.K.Gedroiz' theory, still dominant in Ukrainian soil science, such horizon is formed when exchangeable sodium is replaced by hydrogen cations. Solod soils have somewhat darker and more compacted illuvial horizons which persist in containing exchangeable sodium. Further down the profile, calcium carbonates and soluble salts may be found. Very often the entire soil profile shows distinct signs of reduction processes (gleying). Salt-affected (or halogenic, in the Ukrainian terminology) soils occupy the accumulative land forms like river valleys, saucer-like depressions, pods, etc. Salt accumulation may reach an intensive character in such "basins". The soils are genetically related with each other in evolution (solonchak → solonetz →solod).

29.2. Areas of Salt-Affected Soils in Ukraine

The total area of salt-affected soils in Ukraine is over 5 mln hectares of which 4.5 mln hectares are a farmland including 2.85 mln hectares of plowland. The saline soils occupy an area of 1.71 mln hectares, and sodicity affected and sodic soils – that of 2.82 mln hectares (the data of the State Land Cadastre of Ukraine, 2000, table 97).

The largest areas of saline and sodic soils are located in the Arid Steppe zone and in the northern part of the Crimean Steppe. They also occur in the Left-bank Forest-Steppe, partly in the Northern Steppe, and the Southern Left-bank Polissya zones. Soils with slight extent of salinity dominate among saline soils (78.0%). Among sodicity-affected soils, slightly sodicity-affected species occupy 81.5%, and moderately sodicity-affected – only 14.0% of their total area. In sodicity-affected soil mosaics, those containing 10-30% of sodic (solonetz) soils occupy 57.3%, those containing 30-50% of sodic soils – 20.2%, and those containing over 50% of sodic soils – 20.5% of their total area.

Table 97. Areas of Salt-Affected Soils in Ukraine (data of the State Land Cadastre of Ukraine, 2000)

Soil-Climatic zones	Farmland, thousand hectares					Plowland, thousand hectares				
	Total area	saline		sodic		Total area	saline		sodic	
		area	%	area	%		area	%	area	%
Polissya and Carpathian mountain in province	795777	136.9	1.7	61.1	0.8	55.20	67.2	1.2	38.0	0.7
Forest-Steppe	145189	580.5	4.0	604.9	4.2	12.0126	188.3	1.6	290.5	2.4
Steppe	172228	791.6	4.6	15394	8.9	143576	455.8	3.2	12136	8.5
AR of Crimea	17962	201.0	11.3	619.4	35.0	12598	136.9	10.9	461.3	36.6
Total	414956	17100	4.1	28248	6.8	331502	848.2	2.6	20034	6.0

Thus, salt-affected soils in Ukraine occur in the majority of regions and occupy about 11% of the area of farmland and 8.6% of the area of plowland.

The presence of salts in soils impedes the growth and development of plants. Soil solution acquires great osmotic pressure. The consumption of water by plants becomes impaired, and so do the synthesis of proteins and the growth of plants. The activity of soil microorganisms slows down. Only the plants specially adapted to such conditions can grow and develop on saline soils. Such plants are called halophytes. Plants growing in saline media can somehow increase their internal osmotic concentrations by producing of organic acids or by the uptake of salts. This process is called osmotic adjustment. The effect of salinity on the plant appears primarily to be energy diversion from growth processes in order to maintain the osmotic differential. One of the first processes from which growth energy is diverted is cell elongation. Leaf tissue cells continue to divide but do not elongate. The occurrence of more cells per unit leaf area accounts for a typically dark green color of osmotically stressed plants. The relative growth of plants in the presence of salinity has been termed their *salt tolerance*.

29.3. Salts Effects on Plants

On soils containing 3 and more percent of salts only salsola, bermuda grass, petrisimonia and other halophytes can grow, on those with 2 to 1% of salts – tall wheatgrass, tall fescue, and kermek, and on those with less than 1% of salts – wormwood, wheat grass and some varieties of clover.

The earlier data were summarized by placing plants in one of several salt-tolerance groups. Subsequent listings were in terms of relative plant growth at various saturation-extract EC (electric conductivity) levels (Table 98). Some recent listings have been given instead of EC at the point of initial yield decline, and percent yield decrease per unit increase in salinity beyond this threshold. Most yield data were obtained from uniformly salinized field plots having nearly constant salinity with depth. For most crops, varietal differences are relatively insignificant.

Table 98. Salt Tolerance of Plants (from H.L.Bohn, B.L.McNeal, and G.A.O'Connor. Soil Chemistry. A Wiley Interscience Publication. John Wiley and Sons. 1985. p.248-249)

Crop	EC (ds m^{-1} at 25°C) at which yield will be decreased by		
	10%	25%	50%
Forage group			
Bermuda grass (Cynodon dactylon)	13	16	18
Tall wheatgrass (Agropyron elongatum)	11	15	18
Tall fescue (Festuca arundinacea)	7	10.5	14.5
Barley, hay (Hordeum vulgare)	8	11	13.5
Perennial ryegrass (Lolium perenne)	8	10	13
Beardless wild rye (Elymus triticoides)	4	7	11
Alfalfa (Medicago sativa)	3	5	8
Orchardgrass (Dactylis glomerata)	2.5	4.5	8

Meadow foxtail (Alopecurus pratensis)	2	3.5	6.5
Alsike and red clovers (Trifolium hybridum)	2	2.5	4
Field Crops			
Barley, grain	12	16	18
Sugar beet (Beta vulgaris)	10	13	16
Wheat (Triticum aestivum)	7	10	14
Sorghum (Sorghum vulgare)	6	9	12
Soybean (Glycine max)	5.5	7	9
Rice, paddy (Oryza sativa)	5	6	8
Corn (Zea mays)	5	6	7
Broadbean (Vicia faba)	3.5	4.5	6.5
Flax (Linum asitatissium)	3	4.5	6.5
Field bean (Phaseonlus vulgaris)	1.5	2	3
Vegetable Crops			
Beets (Beta vulgaris)	8	10	12
Spinach (Spinacia deracea)	5.5	7	8
Tomato (Lycopersium esculentum0	4	6.5	8
Cabbage (Brassica oleracea var. capitata)	2.5	4	7
Potato (Solanum tuberosum)	2.5	4	6
Lettuce (Lactuca sativa)	2	3	5
Onion (Allium cepa)	2	3.5	4
Carrot (Daucus carota)	1.5	2.5	4
Greenbean (Phaseoulus vulgaris)	1.5	2	3.5

29.4. Salts Origin and Behavior in Soils

Though one might think that naturally saline areas would be better left unfarmed, the typically favorable year-round climates of many such areas, the desire to develop all portions of a given (privately owned) area due to crop processing needs, and the expense of installing and maintaining a water conveyance system can dictate the development of many naturally saline areas.

The three main natural sources of soil salinity are mineral weathering, atmospheric precipitation, and fossil salts (those remaining from former marine or lacustrine environments). Salts are also added to soil through human activities, including use of irrigation waters and highly saline brines or industrial wastes. The ultimate source of all soil salts are the exposed rocks and minerals of the earth's crust, from which salts have been released during chemical and physical weathering. In humid areas, soluble salts are carried down through the profile by percolating rainwater and ultimately are transported to the ocean or to inland seas. In arid regions, leaching is generally more localized. Salts tend to accumulate because of the relative scarcity of rainfall, high evaporation and transpiration rates, or landlocked topography.

Minerals rarely dissolve congruently (in strict proportion to their composition). They tend to release their most soluble components. A mineral-high in calcium and magnesium, may, therefore, initially release significant amounts of sodium and potassium to the percolating solution as the most soluble impurities.

The so-called fossil salts can introduce large amounts of salinity to even small quantities of soil water. Release of fossil salts to percolating water also contributes materially to the salinity. Fossil salts can also be dissolved when water-storage or water-transmission structures are placed over saline sediments.

Appreciable salt can also be deposited in an area from the atmosphere. Although inland precipitation is relatively salt-free, rain droplets actually form around tiny condensation nuclei consisting of salt or dust particles. The total salt concentration in rainfall water may be as high as 50 to 200 mg L^{-1} near the sea coast, but rapidly decreases to only a few mg L^{-1} in the continental interior. The salts in rain near the seacoast are high in sodium, chloride, and magnesium. Inland precipitation is dominated by calcium and magnesium sulfates and bicarbonates.

A more common occurrence of soil salinity is local salt accumulation. Soil in low-lying areas, even in arid regions, may have high water tables. Water from groundwater tables within a few meters of the surface can move by capillaries to the soil surface, where it evaporates and leaves behind its salts. The soil salinity may be expressed in percentage or in units of electrical conductivity.

Many present-day salt-affected soils result from human activities. Salts may be transported from areas of over-irrigation to accumulate in poorly drained areas. As drainage waters or irrigation return-flows evaporate, high concentrations of salts may remain.

Salts also accumulate in under-irrigated fields, particularly, if relatively saline irrigation waters are used. The salt concentration of the soil solution increases steadily as water is removed during plants growth. Proper irrigation management includes periodic irrigation with water in excess of plant needs, in order to leach accumulated salts from the plant root zone. As the salt concentration of the soil increases, so does the potential for salinity effects on plants growth. Early appraisals of the salinity of irrigation waters were generally carried out in terms of total dissolved solids (TDS), which is still employed in Ukraine for this purpose.

Several empirical relationships have been developed for converting one type of water quality analysis to another. One of them, used for solutions in the EC range from 0.1 to 5 ds m^{-1}, has the following form:

$$TDS(mg \times L^{-1}) = EC(dsm^{-1}) \times 640 \qquad (20.1)$$

For the understanding of certain soil formation processes in halogenic (salt-affected) soils it is necessary to consider the ways of soda formation in soil environments. There are several ways of sodium carbonate (bicarbonate) formation in soils: chemical, colloidal-chemical, and biochemical.

A chemical way may be illustrated by the reactions:

$$CaCO_3 + 2NaCl + H_2O + CO_2 \rightarrow 2NaHCO_3 + CaCl_2;$$

$$CaCO_3 + Na_2SO_4 + H_2O + CO_2 \rightarrow 2NaHCO_3 + CaSO_4$$

The process involving colloidal adsorbent (SAC) may be like this:

$$[SAC^-]Na^+ + H_2CO_3 \rightleftarrows [SAC^-]H^+ + NaHCO_3;$$

$$[SAC^{2-}]2Na^+ + Ca(HCO_3)_2 \rightleftarrows [SAC^{2-}]Ca^{2+} + 2NaHCO_3$$

Sulfate-reducing bacteria may help the folowing processes:

$$Na_2SO_4 + 2C = Na_2S + 2CO_2;$$

$$Na_2S + H_2CO_3 = 2NaHCO_3 + H_2S$$

Sodium bicarbonate can turn into sodium carbonate and undergo a hydrolysis:

$$2NaHCO_3 \rightarrow Na_2CO_3 + CO_2 + H_2O;$$

$$Na_2CO_3 + H_2O \rightarrow NaOH + H_2CO_3$$

$$NaOH \rightleftarrows Na^+ + OH^-$$

Beside sodium sodicity, there is magnesium sodicity with exchangeable magnesium content up to 50% of the soil's CEC.

29.5. Composition and Properties of Salts Occurring in Salt-Affected Soils

The most widespread in salt-affected soils of Ukraine are chlorides, sulfates, and carbonates of calcium, magnesium and sodium which, at certain content in the soil, cause the stunting of plant growth and development, if not the death of plants.

Carbonates (salts of carbonic acid) are widely occurring in soils, parent materials, and ground waters of the Arid Steppe, the Steppe, and the Forest-Steppe zones of Ukraine. Their role in soil salinity and toxicity for plants depends upon their type, amount, and solubility.

Calcium carbonate $(CaCO_3)$ has a very low solubility(0.065 g×L^{-1}). In the presence of carbonic acid its solubility increases to 166 g×L^{-1}, owing to calcium bicarbonate formation:

$$CaCO_3 + H_2CO_3 = Ca(HCO_3)_2.$$

Without carbonic acid, the solutions of $CaCO_3$ may be quite alkaline (pH up to 10-10.2). Carbon dioxide in soil pores lowers the pH to 7.5-8.5. Owing to its low solubility, calcium carbonate is not toxic to most crops. But the plants adapted to acid soils (rye, potatoes, flax, and lupin) develop more slowly and do not give high yields on calcareous soils.

With high water table, $CaCO_3$ accumulates in the soil as a result of evapotranspiration:

$$Ca^{2+} + CO_2 + H_2O \rightarrow CaCO_3 + 2H^+$$

In arid regions, the content of $CaCO_3$ in the soil may reach 10-20% of its weight. Soils with unbroken layers of $CaCO_3$ in their profiles are cemented and impenetrable to water and plant roots.

Magnesium carbonate $(MgCO_3)$ has a greater solubility (1.29 g×L^{-1}) compared to $CaCO_3$. Owing to the formation of magnesium bicarbonate, the solubility increases to 190 g×L^{-1}:

$$MgCO_3 + H_2CO_3 = Mg(HCO_3)_2$$

Hydrolysis results in considerable alkalinity (pH 10-11) of the resulting solution. Free $MgCO_3$ seldom if ever occurs in soil solution since magnesium-ion is adsorbed by the SAC.

Sodium carbonate (Na_2CO_3) is well soluble in water. The solubility of a given salt may reach 178 g×L^{-1} at the temperature over 20°C. Sodium carbonate hydrolysis results in a very alkaline reaction of the soil solution (pH over 12). This salt is very toxic for most plants. It can cause a noticeable dispersion of soil colloids and in this way ruin soil aggregates and make physical properties of the soil very unfavorable. Leaching does not remediate saline soils containing significant amounts of Na_2CO_3, as it may cause the growth of soil sodicity and further soil degradation. Soil productivity can be considerably reduced in the presence of as little as 0.05-0.1% of sodium carbonate.

Sodium bicarbonate $(NaHCO_3)$ is less soluble (99 g×L^{-1}) than sodium carbonate. It is also much less toxic as its hydrolysis yields only a very small amount of hydroxyls. It is formed in a soil solution according to the reaction:

$$Na_2CO_3 + H_2CO_3 = 2NaHCO_3$$

The process is accelerated by CO_2 in soil air. The reverse process is also possible with the reduction of CO_2 content in soil air:

$$2NaHCO_3 \rightarrow Na_2CO_3 + H_2O + CO_2$$

The process is favored by an increase in soil temperature to 28-32°C. The growth of temperature may even cause the growth of irrigation water alkalinity. Some researchers of soil genesis in the Arid Steppe zone point to the role of the mentioned process in sodicity development in chestnut soils. But even such soils have usually a very low if any content of Na_2CO_3 as they contain gypsum which can remediate such a morbid soil condition according to the equation:

$$Na_2CO_3 + CaSO_4 = CaCO_3 + Na_2SO_4$$

Saline and sodic soils may contain up to 5% of sodium carbonates. At the temperature below 8°C the solubility of soda is very much reduced. To add, soils containing it, have a low water penetrability. Owing to these factors, sodium carbonates and bicarbonates tend to accumulate in the profile.

Ground water evaporation leads to a double salt $(Na_2CO_3 \times NaHCO_3 \times 2H_2O)$ crystallization and accumulation in soil horizons. Soil pH may be within 9-10 units.

River and ground waters in the Polissian and Forest-Steppe zones of Ukraine have salt concentration within 0.5-3 g×L^{-1}, with significant content of sodium carbonate. Salt-affected soils in this region have an alkaline reaction.

Potassium carbonate (K_2CO_3) occurs in the soils much more seldom than sodium carbonate. It is well soluble in water (1110 g×L^{-1}), and its properties are like those of sodium carbonate. Sulfates are in various concentrations in nearly all types of soils.

Calcium sulfate $(CaSO_4)$ is sparingly (1.76 g×L^{-1}) soluble in water and contained in soils, parent and underlying materials of the Arid Steppe zone. It is accumulated in the form of gypsum $(CaSO_4 \times 2H_2O)$, which crystallizes in a real variety of forms. It is not physiologically toxic to plants.

The soil layers with abundant gypsum accumulation are cemented by it and, therefore, impenetrable to roots, air, and water. Gypsum is used to amend sodic and saline soils if the latter contain soda.

Magnesium sulfate $(MgSO_4)$ is a typical component of salt-affected soils. Owing to high solubility (700 g×L^{-1}), it is one of most toxic salts for plants. As epsomite $(MgSO_4 \times 7H_2O)$, it is contained in ground water and salt lakes of the southern Ukraine including the north of the Crimean peninsula. It never accumulates in the soil in its pure form only in combination with other soluble salts. Soils containing magnesium sulfate are of low productivity.

Sodium sulfate (Na_2SO_4) is also a typical component of salt-affected soils, groundwater, and salt lakes. Its toxicity is about one-half to one-third that of magnesium sulfate, and its solubility (about 529 g×L^{-1}) increases with temperature. In the warm period of the year, sodium sulfate, alongside the other soluble salts ($MgSO_4$, $MgCl_2$, and $NaCl$), migrates up the profile to the soil surface, whereas in the cold periods it is immobile due to low solubility and is not leached by the downward water flow.

Being precipitated, sodium sulfate forms the transparent crystals of mirabilite $(Na_2SO_4 \times 10H_2O)$. The rise in temperature causes mirabilite to dehydrate and form tenardite (Na_2SO_4). Sometimes, sodium sulfate crystallizes in combination with calcium sulfate, forming glanberite $(CaSO_4 \times Na_2SO_4)$, which accumulates as a fluffy silt on the surface of solonchaks (i. e. very saline soils).

Potassium sulfate (K_2SO_4) is characterized by the properties similar to those of sodium sulfate, but its toxicity is much lower owing to its lower solubility (100 g×L^{-1}).

Chlorides are the salts of hydrochloric acid. They accumulate in the southern latitudes and are widespread in saline soils and salt lakes. Both their solubility (342-745 g×L^{-1}) and toxicity are high. Chlorides cause soil salinity on the Black and the Azov sea coasts as well as on the Sivash.

Calcium chloride $(CaCl_2)$ is seldom found in soils in the presence of Na_2SO_4 and Na_2CO_3. It easily precipitates into $CaSO_4$ and $CaCO_3$. $CaCl_2$ is very toxic for plants though less toxic than $MgCl_2$ and $NaCl$.

Magnesium chloride $(MgCl_2)$ is found in salt-affected soils, ground water and salt lakes – more often than $CaCl_2$. Its solubility is high (up to 1600 g×L^{-1}), and it is very toxic to plants. Magnesium chloride possesses high hygroscopicity and is capable to absorb water vapor from the air even at low temperature. Crystalline precipitates of $MgCl_2$ readily dissolve under such conditions, and soils, containing this salt, remain wet for a long time after rain. Soils with high $MgCl_2$ content are difficult to reclaim.

Sodium chloride $(NaCl)$ is a permanent and ubiquitous component of the salt-affected soils in the south of Ukraine. Its high solubility (357 g×L^{-1}) causes its high toxicity for most plants. Plant growth is seriously impeded when its content reaches 0.1%, whereas the soils with 2-5% of $NaCl$ are barren for crops.

Potassium chloride (KCl), by its properties (solubility about 342 g×L^{-1}), is very close to sodium chloride, but it is seldom found in saline soils in considerable quantities. At high concentrations its toxicity may be nearly like that of $NaCl$.

Nitrates and borates also belong to soluble salts, the excess of which may be toxic or cause the formation of their toxic derivatives like nitrites and nitrosamines.

By toxic effects on plants, the soluble salts form the following series in the order of their toxicity diminishment:

$$Na_2CO_3 \rightarrow NaCl \rightarrow MgCl_2 \rightarrow CaCl_2 \rightarrow NaHCO_3 \rightarrow MgSO_4 \rightarrow Na_2SO_4$$

The capacity of soluble salts to migrate in soil profile depends on their solubility. The greater the solubility, the faster and to larger distances the salts migrate. Carbonates of calcium and magnesium quickly form saturated solutions, precipitate, and accumulate in parent materials and soils. They cannot migrate to large distances. Gypsum has a slightly higher migration ability, especially in combination with more soluble salts. The presence of chlorides in solution increases the solubility of gypsum from 1.76 to 10 and even 15 g×L^{-1}). The solubility of gypsum is 27 times that of calcium carbonate and the zones of their accumulation are, therefore, separated from each other. But, in many cases, $CaSO_4$ and $CaCO_3$ accumulate in the same soils (chernozems, chestnut, and sodic soils). Magnesium and calcium chlorides, as well as sodium, potassium, and calcium nitrates have a very high migration capacity. They may accumulate at great distances from the places of their origin. Sodium chloride solubility changes little with the temperature of soil environment. So, its migration may be quite considerable in both cold and warm climates.

The salts of sedimentary rocks migrate with water, and accumulate in low areas of topography. Part of water is lost to evaporation and infiltration, whereas the salts, having reached their solubility products, precipitate (crystallize) in deluvium, proluvium, cones of discharge, deltas, and on the terraces. That is the way for the hydromorphic accumulation of salts in parent materials and soils.

Sodium carbonate and sulfate are highly soluble at 20-30°C. Their solubility rapidly drops at 10°C and becomes very low at 0°C and below it. This causes rapid oversaturation of ground water and soil solution with sodium sulfates and carbonates which crystallize and are separated from chlorides. Vegetation also effects salt redistribution in a soil profile. Owing to selective absorbing capacity, the plants assimilate necessary cations and anions from the lower soil horizons. When plant residues decompose, the salts accumulate in the layer of the soil. This causes the changes in salinity of upper and lower soil horizons.

29.6. Solonchaks (Very Saline Soils)

On most continents, except Europe, the dominant types of salt-affected soils are the saline soils ones. Considering the recent scientific achievents and the amount of available data, it may be stated, after I.Szabolcs (1974), that the ratio between saline and alkaline (sodic) soils in Europe indicates the prevalence of the latter. The total extent of saline soils here constitutes approximately less than a quarter of all the area covered by salt-affected soils.

Solonchaks are the soils containing large amounts of soluble salts (over toxicity threshold) in the upper horizons of a soil profile. In Ukraine solonchaks are spread locally on the terraces of the rivers (the Dnipro, the southern Bug, the Dnister, the

Danube, etc) and on the coasts of the Black and the Azov seas. Solonchaks do not have specific horizons in their profiles. They have relic horizons preserved from the former zonal soils and considerably transformed by the effects of salinity. They may be represented by deluvial sediments pierced by abundant veins of salts or by the coastal silt.

The profile of a solonchak is morphologically indistinct. Soluble salts are found in the form of whitish spots, crusts, crystals, etc. By morphologic features, it is usually possible to identify a saline horizon of OM accumulation (H_s), transitional saline horizons (H_{ps} and P_{hs}), and saline and frequently gleyed parent materials (P_{gls}). Signs of reduction processes in the lower portion of a soil profile may be in the form of ochric, rustic, and black points or dull grey spots.

29.6.1. Solonchaks Classification

The Ukrainian soil classification (M.I.Polupan, 1988) does not identify solonchaks as a soil type. Soil salinity is taken into account while determining soil genus, whereas type and extent of salinity determine soil species. International soil science (Russia, USA, and the countries of Western Europe) considers solonchaks as a soil group. The Soil Map of the World (1982) identifies 27 soil groups, including solonchaks.

According to the “Classification and Diagnostics of Soils in the USSR’ (1972), solonchaks are divided into two types: **automorphic** and **hydromorphic** (Table 99).

Table 99. Classification of Solonchaks (Dokuchayev Soil Science Institute, 1977)

Subtype	Genus	Species
Type: Hydromorphic (saline groundwater within 0.5-3 m of soil surface)		
Typical Meadow Swampy Coastal Secondary	By the type of salinity (soil extract analysis)	By morphology of the surface horizon: crusty, pudgy, wet, black by the character of salt distribution in profile: surface or deep-profiled
Subtype	Genus	Species
Type: Automorphic (ground water table over 10 m deep)		

Typical Takyr-like	By the type of salinity; by the source of salinity; lithogenic, ancient hydromorphic, biogenic	Determined like in hydromorphic type

The following subtypes are identified among hydromorphic solonchaks:

1) **Typical solonchaks** are formed with a high water table (within 2-4 m of soil surface) of very saline ground water. Their “solonchaks” properties are expressed most distinctly. The vegetation on such soils is scanty or absent at all. The soil profile is very slightly differentiated: only the surface horizon is expressed clearly and contains abundant salt accumulations.

2) **Meadow solonchaks** are formed under scanty meadow vegetation as a result of meadow soil salting. Meadow soil profile is preserved enough to show the horizons: $H_{dgls} + H_{pgls} + P_{hgls} + P_{gls}$. Meadow solonchaks usually contain more OM than the other subtypes.

3) **Swampy solonchaks** are formed with soluble salt accumulation in swamp soils. The vegetation here consists of salsola, camphorosma, sedges, reeds, and other species. The signs of gleying are abundant all over the profile. There may also be a surface horizon with bits of peat. Mineralized ground water table is within 1-0.5m of soil surface.

4) **Marine solonchaks** are formed in the youngest of marine deposits. They are covered by a pudgy crust of salts, under which there is a sandy or loam-sandy horizon with abundant shells and high content of chlorides. The profile is gleyed from some shallow depth.

5) **Secondary solonchaks** are the result of a faulty irrigation. In most cases, it is the lack of a proper drainage. As the saline ground water table becomes higher, so does the content of soluble salts in a soil profile.

The following subtypes are identified among the automorphic solonchaks:

1) **Typical,** which are characterized by the greatest accumulation of soluble salts on the soil surface.

2) **Takyr-like solonchaks,** which show the first stage of typical solonchak leaching. A thin (1-2 cm) crust of leached material forms on the surface.

The source of soil salinity is taken into account in determining solonchak genuses. By this criterion the following genuses are identified:

1) **Lithogenic** – formed in saline parent materials;

2) **Ancient (relic) hydromorphic** – formed from hydromorphic solonchaks which lost connection with ground water as water table became lower.

3) **Biogenic** – formed as a result of biologic fixation of nitrogen by specific microorganisms and accumulation of nitrates.

Type of salinity imparts some specific features to the surface horizons of solonchaks:

1) **Crusty** surface is typical of sodium chloride $(NaCl)$ salinity. The crust of salt on the surface is rather hard.

2) **Pudgy** solonchaks contain much of sodium sulfate $(Na_2SO_4 \cdot 10H_2O)$. The soil surface horizon is very dry and fluffy, with bulk density not over 1.0 g/cm^3.

3) **Wet** solonchaks contain hygroscopic salts ($CaCl_2$ and $MgCl_2$), and their surface layer is, therefore, dark and wet to touch.

4) **Black** solonchaks, characterized by a high content of sodium carbonate (Na_2CO_3) which increases the solubility of SOM. After rain or irrigation, puddles are formed on the surface and water does not penetrate deeply into the soil, being evaporated from the surface. The soil surface becomes black in color.

The most widespread in Ukraine are sulfate-sodic (Near-the-Dnipro region and the terraces of the Dnipro valley), chloride-sulfatic, and sulfate chloridic (the Black Sea and the Sivash shores regions) solonchaks.

It is important to consider the depth of soil salinity by which the saline soils may be solonchaky (soluble salts with 0-30 cm of depth), highly solonchaky (30-50 cm), moderately deeply solonchaky (50-100 cm), very deeply solonchaky (100-150 cm), and nonsolonchaky or deeply saline (150-200 cm).

Besides solonchaks, there are slightly, moderately, and very saline soils in Ukraine, which preserve their zonal features to the greater extent than solonchaks.

29.6.2. Secondary Salinity of Soils

The secondary salinity of soils is caused by anthropogenic agents, such as irrigation. In a broader frame of references, secondary salinity may develop:

1) under the influence of saline ground water;
2) as a result of irrigation water salinity;
3) under the influence of salt-containing deep horizons of soils and parent (or/and underlying) materials.

The secondary salinity caused by saline ground water can occur when the water rises up the capillaries to the upper horizons. The penetration of salts into the soil by a capillary rise can cause the development of a secondary salinity under the influence of irrational (excessive) irrigation. Such kind of salinity may also develop in soils surrounding irrigated areas or around the fish-breeding ponds. To prevent the capillary rise of saline ground water to soil horizons, the ground water table should be lowered below critical depth.

In addition to the general osmotic effects, many plants are tensitive to specific ions in irrigation waters or soil solution. Boron toxicity is probably most common. Boron is more difficult to control than is salinity in general.

Direct sensitiving to exchangeable or soluble sodium is more apparent at low salt levels, and therefore is difficult to differentiate from the effects of sodium on soil permeability. For plants that are extremely sensitive to sodium, as little as 5% of exchangeable sodium may lead to toxic accumulations of sodium in leaf tissues. Some grape rootstocks exhibit up to thirty-fold differences in their abilities to exclude chloride ions. The selection of a rootstock that screen out ions may prevent toxic accumulations in plant tops.

Another mechanism for potential salt injury to plants involves **nutritional imbalances.** An example is the bicarbonate toxicities reported for some saline environments. These result primarily from a reduced iron availability the high pH levels common to high bicarbonate soils.

High sodium levels can conceivably lead to calcium and magnesium differences. High soil pH levels also might lead to high concentrations of soluble aluminum [$Al(OH)_4^-$ species]. Many of the suppositions are still largely speculative.

29.6.3. Removing Soluble Salts from Saline Soil Profiles

The essential and most feasible way to remove the salts from the soil is the leaching of the latter to a sufficient depth with the employment of horizontal, vertical or combined drainage. Other techniques include the following:

1) **Mechanical removal.** The salts, or rather soil crusts, are scooped up and moved beyond the limits of an irrigation area. Such operation is useful before leaching.

2) **Plowing down of salts** is employed when lower soil horizons are not saline and salts are concentrated in a surface horizon, but even in them the salinity is not too high. In this way the salts may be “diluted” in a plow layer to concentrations, not toxic for crops.

3) **Surface leaching** is employed to remove the salts from the root-abundant horizons of heavy-textured soils with a low infiltration rate. Considerable volumes of water (up to 20-30 thousand m^3/ha) are applied to the soil divided in portions with a certain pattern in time. Very often surface leaching is combined with the cultivation of rice or fish breeding on irrigated areas.

The management of salt-affected soils should be centered around the maintenance of a salt balance for a given field or irrigation project. The quantity of salt leaving the area should be greater than or equal to, the quantity of salt entering the area if long-term agriculture is to be maintained.

To calculate the leaching requirements (LR), one needs an estimate of the allowable EC of the saturated extract (or allowable salt concentration) such as can be obtained from existing salt tolerance data. Leaching rate is the amount of water which should be applied to the soil during the period of leaching. Leaching is carried out preferably in autumn or in winter, when the evaporation is limited, in two stages. At the first stage, the soil is saturated with water to FC to dissolve the salts. At the second stage, the salts are removed from the soil by an additional portion of leaching water.

In the practice of farmland management, it is possible and necessary to obtain most desirable effects of leaching using a minimal amount of water. For this purpose, the area of leaching must be specially prepared. It must be leveled, deeply plowed, harrowed and divided into cell areas with earthen walls between them.

L.P. Rozov proposed to compute the LR by the formula:

$$LR = FC - SW + nFC \qquad (20.2)$$

where: FC = the amount of soil water (m^3/ha) at field capacity; SW = the amount of water in the soil (m^3/ha) before the leaching practice, and its dependence on the extent of soil salinity and dimensionless.

According to L.P.Rozov, if the soil is saturated to FC (field capacity), then a single (n = 1) additional portion of leaching water may remove 60 to 90% of the initial amount of salts from the soil.

According to the estimates of the American scientists (H.Bohn, B.McNeal, and G.O'Connor, 1985), the passage of 1 m leaching water m^{-1} soil depth under ponding conditions normally removes approximately 80% of the soluble salt from soils. But these scientists also think that leaching under unsaturated conditions, such as with the use of intermittent ponding or sprinkler irrigation, may lower this quantity of water as much as three to five times. Boron removal can require up to three times more water, because boron is retained to some extent by soils.

V.R.Volobuyev proposed his equation for the computation of LR, which is in a widely used in the former USSR countries:

$$LR = 10000 \times h \times \alpha \log S_f / S_t \qquad (20.3)$$

where LR is expressed in cubic meters per hectare, h is the depth of leaching in meters, α is a dimensionless parameter which depends on soil texture and type of salinity, S_f and S_t are respectively the factual and tolerable percentage of soluble salts in a layer of leaching (% by weight of oven-dry soil). Table 100 gives the values of α.

Table 100. Values of α-parameter used in V.R. Volobuyev's formula

Soil texture	Type of Soil Salinity			
	chloridic	sulfate-chloridic	chloride-sulfatic	sulfatic
Sand, loamy sand	0.62	0.72	0.82	1.18
Loam	0.92	1.02	1.12	1.41
Light clay	1.22	1.32	1.42	1.78
Medium clay	1.80	1.90	2.10	2.40
Heavy clay	2.70	2.80	3.00	3.80

29.7. Solonetz (Sodic, Alkali) Soils

According to I. Szabolcs, alkali soils are affected by sodium salts capable of alkaline hydrolysis (mainly $NaHCO_3$, Na_2CO_3, and Na_2SiO_3). According to the present-day American Classification of Salt-Affected Soils (1973), sodic soils have exchangeable sodium percentage (ESP) over 15% or exchangeable sodium ratio (ESR) over 0.15.

According to a well-known Soil Science Textbook, edited by I.S.Kaurichev (1989), “Solonetz soils are the soils containing in adsorbed state a great quantity of exchangeable sodium and sometimes magnesium in illuvial (solonetzic) horizon”.

Sodic soils present a particularly difficult management problem. The permeability of these soils to water is generally so low, that water passes through them only slowly. The pH of soda (black alkali) solonetz soils is commonly greater than 9 or 9.5, and the clay and organic fractions are dispersed.

29.7.1. Solonetz Soils Genesis and Classification

According to K.K.Gedroiz, solonetz soils are saline soils derivatives. Sodium from some soluble salts (Na_2CO_3, $NaHCO_3$, etc) gets into the soil adsorbing complex (SAC), dispersing a part of soil colloidal matter. This dispersing action of exchangeable sodium is possible under low concentrations of soluble salts. Salt-affected soil evolution procedes in the following sequence:

$$Sodium \quad solonchak \rightarrow solonetz \rightarrow solot \quad (solod)$$

Partial dispersion of soil colloids allows them to be leached down the profile from the eluvial to illuvial (solonetzic) horizon. Dispersed organic matter may accumulate at the surface of poorly drained areas and impart a black color, hence the common name “black alkali”. Solonetz soils often occur in small patches, or slick spots, less than 0.5 hectare in extent. Such patches occupy slight micro-relief depressions, which become unsuitable for plants growth. For a long time, exchangeable sodium has been considered the main cause of solonetz formation. Such a view dominates to this day but in some typical solonetz profiles with less than 10% of the CEC occupied by sodium, up to 40-60% of it was found to be occupied by magnesium. In many cases, magnesium was thought to be similar to calcium in its action upon the soil, and solonetz soils with low exchangeable Na^+ content were thought to be “residual” or “relic”.

At present, a new theory was suggested (Mikhailichenko, 1979) on the role of hydrophillic colloids in the origin and development of solonetz soils. Soluble colloidal plasm is formed only in the presence of hydrolysable sodium salts, whereas magnesium solonetz soils are found in nature only owing to the inability of Mg^{2+} to convert hydrophyllic colloidal plasm into precipitated gel, though Ca^{2+} is capable of performing such a convertion. Various hypotheses continue to arise. Solonetz soils classification is complicated and illustrated by Table 101.

Table 101. Solonetz Soil Classification

Type	Subtype	Genus	Species
Steppe	Chernozemic Zonal	By depth of salinity: solonchaky (0-30 cm);	By depth of solonetz horizon (H_i) placement: crusty (1-2 cm); shallow (3-5 cm); medium (5-15 cm); deep (> 15 cm)
Meadow-Steppe	Chestnut	highly solonchaky (30-50 cm);	By exchangeable Na^+ content in the HI horizon: low-sodium (< 10%); moderately-sodium (10-25%); highly-sodium (> 25 %).
Meadow	Meadow-chernozemic	deeply solonchaky (50-100 cm)	By the structure of the HI horizon: Columnar, Nutty, Prismatic, Blocky.
Lithogenic	Meadow-chestnut	very deeply solonchaky (100-150 cm);	By the extent of solodizity: Slightly (no E_{gl}),

			Moderately ($E_{gl} < HE_{gl}$), and strongly ($E_{gl} > HE_{gl}$) solodized.
Chernozem-like	Chernozemic-meadow	Nonsolonchaky (deeply saline, 150-200 cm); By the extent of salinity: very saline, moderately saline, slightly saline;	
	Azonal Lithogenic Chernozemic Modal solodized	By the type of salinity soda, soda-sulfatic, soda-chloridic, etc. By the depth of gypsum and carbonate occurrence: highly carbonatic (< 40 cm); deeply carbonatic (> 40 cm); highly gypsiferous (< 40 cm); deeply gypsiferous (> 40 cm);	

29.7.2. Solonetz Soils Remediation and Use

The aim of any remediation of solonetz soils is the replacement of exchangeable sodium from the SAC and its substitution by exchangeable calcium. This improves agrophysical and chemical properties of sodic soils and increases their productivity.

Finally ground gypsum $(CaSO_4 \cdot 2H_2O)$ is broadcast and mixed with the soil. The amount of gypsum needed to replace the exchangeable sodium is the **gypsum requirements (GR).**

All this pertains to chemical remediation of solonetz soils. The up-to-date approach to the improvement of sodicity-affected soils in Ukraine is a differentiated system of practices. Sodicity-affected soils needing improvement are divided into 5 groups (or categories). **The first group** includes the most potentially productive soils, like sodicity-affected southern chernozems and dark-chestnut soils in complexes with sodic soils (up to 10% of total area). This category does not require profound practices of soil improvement. They need proper fertilizing, water conservation practices, proper irrigation, and the selection of sodicity-resistant crops. The soils may be used in the field and forage crop rotations.

The lands of the second category have a good self-improving ability when subjected to deep plowing. These are meadow-chestnut and dark-chestnut sodicity-affected soils in combination (up to 25%) with sodic areas and not too deep line of carbonates (up to 40-50 cm of soil surface). **The third and fourth categories need the application of gypsum.** They may be in the field crop rotations, but more often than not it is advisable to turn such areas to grasslands and pastures. **The fifth category demands phytomelioration** and use in grasslands and pastures. It does not pay to amend them with gypsum.

The profile of a chestnut solonetz soil of the Askania-Nova reserve (Kherson region), photographed and described in the *Soil Atlas of the Ukrainian SSR,* edited by M.K.Krupsky and M.I.Polupan (1979) has the following horizons:

$HE_{d\ 0\text{-}7\ cm}$: Humuso-eluvial horizon; chestnut-grey; heavy loam; cloddy with powder; slightly platy (indistinct horizontal layering); loose; abundant roots; boundary clear.

$E_{h\ 8\text{-}15\ cm}$: Eluvial horizon; light grey; heavy loam; platy; abundant roots; boundary very distinct (sharp).

$I_{h\ 16\text{-}24\ cm}$: Illuvial (solonetzic) horizon; dark-chestnut; medium clay; columnar-prismatic; compacted; with thin cracks; scanty roots; transition gradual.

Pl$_{h/k}$ $_{25\text{-}45\ cm}$: transition horizon; illuviated; chestnut with brown tint; light clay; nutty-prismatic; compacted; from the depth of 40 cm shows effervescence with dilute HCl; scanty roots; transition gradual.

P$_k$ (h) $_{46\text{-}55\ cm}$; loess, slightly dirtied with OM.

P$_{ks}$ $_{56\text{-}170\ cm}$; loess, straw-colored, light clay, carbonate nodules, crystals of salts, abundant gypsum.

When solonetz spots exceed 25% in area, the lands should be turned to grasslands. Their productivity can be improved by "phytomelioration", which is the selection of sodicity-resistant plant species. Well developed vegetation excludes physical evaporation from soil surface, lowers ground water table, and reduces the capillary rise of soluble salts to the overlying horizons. Root exudates have acid reaction and this partly neutralizes excessive soil alkalinity. Phytomeliorating plants reduce the content of exchangeable sodium and activate the microbiologic processes owing to soil enrichment with organic matter.

Gypsum is applied in sodicity-affected chernozems if ESP exceeds 5 and in sodicity-affected dark-chestnut soils – when it exceeds. Gypsum rates are computed according to the following formula:

for neutral soils –

$$GR = 0.086(Na - 0.05CEC) \bullet d \bullet h \qquad (20.4)$$

1) for slightly alkaline soils –

$$GR = 0.086(Na - 0.05CEC) + (S - 0.7) \bullet d \bullet h \qquad (20.5)$$

for alkaline (sodic) soils –

$$GR = 0.086[(Na - 0.05CEC) + (S - 0.7)] \bullet d \bullet h \qquad (20.6)$$

and for magnesium (low-sodium) solonetz soils –

$$GR = 0.086[(Na - 0.1CEC) + (Mg - 0.3CEC)] \bullet d \bullet h \qquad (20.7)$$

where GR is expressed in mt/ha of $CaSO_4 \cdot 2H_2O$,

Na – the content of exchangeable sodium in milliequivalents per 100 g of oven-dry soil;

CEC – the cation exchange capacity of a soil in milliequivalents per 100 g of oven-dry soil; d – the bulk density of the soil in g/cm^3; h – the thickness (cm) of a soil layer which needs amendment; S – the content of toxic alkalinity determined by soil extract analysis $(HCO_3^- - Ca^{2+})$, expressed in milliequivalents per 100 g of oven-dry soil; and C = sodium carbonate content $(Co_3^{2-} - Na^{+})$, determined by the soil extract analysis and expressed in the same units as the previous characteristic.

Gypsum reactions in a sodic soil environment can be illustrated by the following equations:

$$Na_2CO_3 + CaSO_4 \cdot 2H_2O \rightleftarrows Na_2SO_4 + CaCO_3 + H_2O$$

$$[SAC^{2-}]2Na^{+} + CaSO_4 \cdot 2H_2O \rightleftarrows [SAC^{2-}]Ca^{2+} + Na_2SO_4$$

$$2NaHCO_3 + CaSO_4 \cdot 2H_2O \rightleftarrows Ca(HCO_3)_2 + Na_2SO_4 + 2H_2O$$

In *Soil Chemistry,* the textbook by H. Bohn, et al (1985) stressed is the fact that the tests of soil exchangeable sodium levels should be made every two to three years to estimate the need for reapplication of gypsum.

The authors of the textbook also point out to the other common acidifying amendments for the reclamation of **calcareous sodic soils**, including sulfuric acid and elemental sulfur. Another source of soluble calcium is lime, if a soil is only slightly sodic and light-textured. Deep plowing to the 0.7-0.9 meter depth has also proved helpful in redistributing subsurface lime and in opening the soil in order to maintain adequate water permeability during reclamation. In most cases, however, lime is not sufficiently soluble to serve as an amendment for sodic-soil reclamation.

29.8. Solod (Solot) Soils

In Eurasia solod and solodized soils occupy only 0.5% of the total area of soils of which solods per se occupy only 0.1%. They occur in the Forest Steppe and the Chernozemic Steppe zones as well as among the soils of arid and semidesert steppes. Everywhere these soils occupy the locked depressions of topography.

29.8.1. Solod Soils Morphologic Features, Genesis, and Properties

Soil profile is very distinctly differentiated into eluvial and illuvial horizons. The soils were formed under conditions of periodic excessive wetness and percolation of the entire profile. Labile products of gleying are redistributed over the profile differentiating it into the eluvial (solodized, solod) and illuvial horizons. Any soil, having such a differentiation on the general “background” of gleying, is solodized to a certain extent, or it turns into a solod proper. According to the theory, advanced by K.K.Gedroitz, the solods are formed after percolation and leaching (desalinization) of sodic soils by water accumulated in the saucer-like depressions. K.K.Gedroitz believed that exchangeable sodium in solod formation was substituted by the hydrogen-ion. In an alkaline environment, developed as the sodium interacted with carbonic acid, the soil adsorbing complex was subject to severe disintegration. One of the morphologic features of such soils is the abundance of amorphous silica solved in 5% KOH in the eluvial portion of a soil profile. Aluminosilicates, subjected to destruction, release much of amorphous silica. Some authors attach great importance in this process to diatomic algae and other organisms.

Chemical processes leading to the release of free SiO_2 may take place during the desalinization of solonetz soils and under the influence of periodic soil leaching with a weak solution of sodium salts. In the latter case, the soil profile is at first subject to sodicity and, subsequently, to “solodisity”, as the percolating waters carry away the products of alkaline hydrolysis. According to S.P.Yarkov and I.S.Kaurichev, the soil anaerobiosis which developes under excessive wetness, instigates the formation of active organic acids (fulvic and low-molecular-weight carbonic acids) and soluble forms of iron, manganese, and other elements which form soluble organic-mineral complexes moving downward in a solod profile formation.

Meadow-chernozemic solods formed in loess consist of humuso-solodized horizon **(H_e):** dark grey with amorphous SiO_2 dusting, cloddy-platy or scaly, humuso-eluvial (solod) horizon **(HE_{gl})** with abundant amorphous SiO_2 and rusty and olive mottling and hard bean-like concretions, illuvial horizon **(I_{ehgl}):** greish brown, nutty, with abundant spots and mottles of gleying, and parent material **(P_{gl}),** though more often noncarbonated, but occasionally containing large nodules of carbonate concretions. Meadow solods form in the areas of higher watertable. Their profile is shorter and more gleyed.

A sharp profile differentiation is evident from the soil texture: eluvial horizon is much poorer in clay than the illuvial horizon underlying it. The chemical composition of the mineral part of the soil also shows a higher SiO_2 content in eluvial and a higher R_2O_3 content in illuvial horizons (Table 102).

The soil extract analysis usually shows a very small amount of soluble salts in solods.

Table 102. Chemical composition and Physico-Chemical Properties of Solod Soils (data of N.I. Bazilevich)

Horizons and their depth (cm)	SOM, %	pH_{KCl}	Chemical composition, % of dry and decarbonated mineral part			Exchangeable cations, milliequivalents per 100 g of oven-dry soil			
			SiO_2	Al_2O_3	Fe_2O_3	Ca^{2+}	Mg^{2+}	Na^{+}	S
$H_{e\ 0-5}$	8.8	4.7	71.7	12.0	2.9	17	5	1	23
$E_{h\ 10-16}$	0.8	3.7	80.4	11.8	2.9	5	3	traces	8
$E_{h\ 17-20}$	0.8	4.0	79.6	11.8	2.8	6	3	1	10
$I_{(h)\ 20-25}$	1.3	5.0	69.4	16.1	5.8	16	13	2	11
$IP_{\ 47-57}$	-	6.6	65.0	17.8	6.8	13			
$P_{\ 120-130}$	-	8.0	74.2	15.1	6.0	-	-	-	-

29.8.2. Solod Soils Classification and Diagnostics

Depending on soil formation conditions, solod soils are divided into three subtypes: 1) forest (typical) solods; 2) meadow (soddy) solods, and 3) meadow-swampy solods.

Typical forest solods develop under birch and aspen-birch forests (“kolkas”) in Ukrainian with locked canopies and slightly developed grass vegetation. Their profiles contain a well-expressed E-horizon. HE-horizon is usually not more than 5 cm in thickness. **Meadow (Soddy) Solods** form is less thick “kolkas” or on the lowland areas with well developed grassland vegetation. A well-developed soddy (H_{ed}, HE_d) horizon is underlain by a solod (E_h, $E_{(h)}$) eluvial horizon.

Meadow-swampy or peat-covered solods occur in lowland areas where they develop under meadow-swampy vegetation including shrubs (willows) with ground watertable about 1 meter below the soil surface. The soddy horizon contains at least the bits of peat. The signs of gleying are well expressed all over the profile.

Solod soils are divided into genuses by the relic features of sodicity and salinity. Solods may be noncarbonatic, nonsaline, nonsodic, sodic, and solonchaky. Meadow and typical solods are sometimes divided into species according to the extent of gleying. By the content of OM in the surface horizon, the solods are divided into the species of light (< 3% of SOM), grey (3-6%), and dark (>6%). Forest solods, as a rule, are nonsaline and nonsodic.

29.8.3. Solod Soils Agricultural Use

Solods are the soils of low natural productivity. Solodized horizons contain little SOM and few nutrition elements. To employ the solods as a farmland, it is, primarily, necessary to increase their OM content by applying sufficient rates of organic manures. Many solods have an acid reaction in the surface horizons and need regular liming. The physical properties of solods are also unfavorable and they need deep subsoiling. There were some not always successful attempts to cover such soils with a layer of more fertile soil material. It pays to do that only when the spots of solods are in the way of soil tillage on large areas of plowland. In the majority of cases, the solod soils should better be left under forest vegetation, which, in a way, may fulfill a wind-breaking function in the field. In arid steppe zone, the gleyed solods in large depression (“pods”) remain flooded for a long period of time. The good side of it is some increase in air humidity which is very important for a given region, even in the prevention of severe wind erosion.

Questions & Assignments

1. Speak on the distribution of salt-affected soils in Ukraine.
2. Define solonchaky and solonetz soils.
3. Speak on the sources of soluble salts in soils.
4. Speak on the effects of soluble salts on plants.
5. Speak on the salt tolerance of field crops.
6. How is soda formed in soil environments?
7. Speak on the capacity of soluble salts to migrate within a soil profile.
8. Which types and subtypes of solonchaks do you know?
9. What is secondary alkalinity and what it is caused by?
10. Speak on the practices of salt-affected soils reclamation.

Chapter 30

Soils of the Carpathian Mountain Province and the Crimean Mountain Region

Mountain soils are formed and distributed in mountain regions according to the law of **vertical zonality**, formulated by V.V.Dokuchaev. Vertical zonality may be defined as the change in the character of soil cover with altitude caused by changes of climate and vegetation.

Soil zones in the mountain systems form the belts within certain ranges of altitude. But the order in geographic distribution of soils formation factors may be violated in some cases of large systems of vertical soil zones due to the phenomena of **vertical zone inversion, migration, and interference.**

Inversion means the opposite order of zones sequence, **migration** is the penetration of one zone into another, and **interference** is the absence of certain zones in their expected sequence.

All factors of soils formation in the mountains differ from their plainland manifestations.

30.1. Soils of the Carpathian Mountain Province

30.1.1. The General Features of Soil Formation

The Ukrainian eastern Carpathian mauntains are situated in the south-western part of the country, stretching out to over 200 kilometers in length and about 100 kilometers in width. Their area exceeds 30 thousand square kilometers. According to A.M. Turenko (1980), the Carpathian Mountain Province is an integrated landform system (Fig. 97), including as its integral parts, the mountain ranges, The Transcarpathian lowland, and the Carpathian plain land.

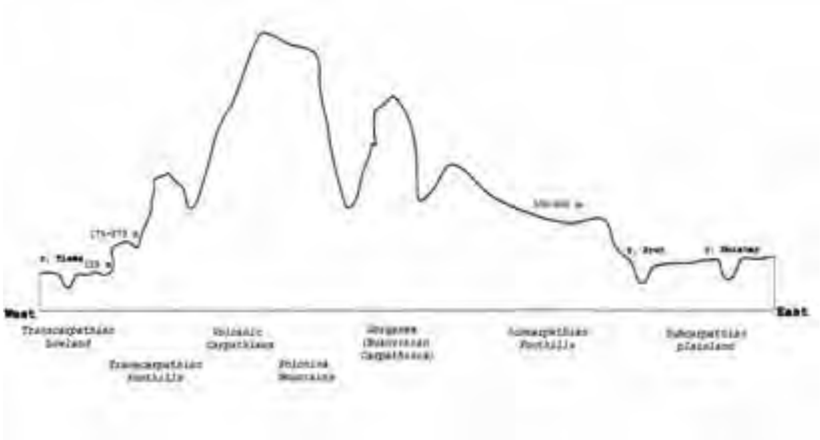


Fig 97. Schematic Transsection Across Ukrainian Carpathians (A.M. Turenko, 1980)

The area of the Ukrainian Carpathians is known for its specific land forms and clearly distinguished vertical zonality of climate, vegetation, and soils.

The Carpathians parent materials are mainly represented by the products of weathering of solid sedimentary, igneous, and metamorphic rocks. Sandstones and slates, which by alternating layering form the so-called Carpathian "flisch", are the most widespread parent materials. In the southern part of the province, the soils develop in the cortex of weathering of andesites, basalts, and other volcanic rocks redeposited in the form of alluvium and diluvium, rich in rubble and stones, the amount of which increases with elevation. Alluvial and deluvial deposits cover sandstones, slates, marls, and other solid rocks.

Noncalcareous loess-like deluvial sediments dominate in the Subcarpathian part of the province. Alluvial loams, underlain by sands and pebbles, are widespread on young terraces of river valleys. The recent deposits of sands, loamy sands, and loams, underlain by the layers of pebbles, are very common on the floodplains of river valleys.

Transcarpathian soils were formed in eluvial and deluvial sediments of heavy texture (clay loams), free from stones and rubble, and in alluvial deposits and loams covering sands and pebbles.

Climate of the Carpathian Mountains Province is formed under the influence of the Atlantic and the continental air masses. It is characterized by high humidity coefficients, exceeding unity, and well defined specific features depending on the absolute altitude and direction of the mountain range. The South-western slope and the Transcarpathian lowland are to a greater extent subject to the impact of warm and humid Atlantic air masses than the north-eastern slope and the Sub-Carpathian territory being under the influence of East-European and Asian continental.

In general, the climate in the Carpathians is humid or excessively humid. By thermal regime and hydrothermal conditions of soil formation, several vertical zones were identified within the area of the Ukrainian Carpathians and adjoining plainland areas. (Table 103).

Changes in climate, depending on the altitude, are the main reason for the vertical zonality of vegetation. Broad-leaved forests consisting of oak, beech, and hornbeam with an admixture of spruce are common on the foothill plains in the Subcarpathians. Low mountain areas are occupied by mixed deciduous forests of oak, hornbeam, and beech, whereas moderately high mountains are covered mainly by mixed broadleave-aciduaous forests rich in spruce. On high-perched meadows called *poloninas* in Ukrainian, meadow grasses are the most common vegetation, but occasionally one can come across the undergrowth of bushes, and thinned-out alder and pine forests with stunted and deformed trees. On the Transcarpathian foothills, oak forests dominate, with hornbeam coppices occupying some elevated hills.

Table 103. Hydrothermal Regime Characteristics of the Ukrainian Carpathians Vertical Zones (I.M.Gogolev, 1986)

Vertical zone, region, and belt	Characteristics in the yearly cycle		
	$\sum_t > 10$°C	Precipitation, mm	Humidity coefficient
Transcarpathian lowland, warm, humid	2800-3000	700	1.0-1.3
Transcarpathian foothills, warm, excessively humid	2600-2800	800	1.3
Subcarpathian, moderately warm, excessively humid	2200-2500	700-800	> 1.3
Warm, humid	2500-2800	600-700	1.0-1.1
Mountain forest, excessively humid	1800-2200	800-1000	> 1.3
Low-mountain, moderate (450-470 m)	1400-1800	1000	>> - >>
Moderately-Mountain, cool (750-950 m)	1000-1400	1200	>> - >>
Moderately cold (650-1500 m)	600-1000	1200-1600	>> - >>
Cold (over 1500 m)	< 600	1200-1600	>> - >>

Vertical changes in climate, vegetation, parent materials, and topography in the Carpathian mountains province caused the changes in soil formation. The brown forest soil formation proved most widespread and intensive here. The development of such process acquires most typical features under the canopy of forest vegetation on the slopes less than 20 degrees in steepness.

Such conditions complemented by the subsoil runoff of water favor the formation of aluminum iron-humus complexes which, moving in the soil along the sloping subsurface horizons, accumulate mainly in the surface horizons of a soil profile. The lateral type of migration of the dissolved and suspended material creates in the soil profile an infiltrative transit of organic compounds, iron, and finely dispersed particles.

Brown forest soil formation. Any compaction or reduction of soil infiltrability reduces the influx of soluble materials to the deeper horizons. Organic-mineral complexes are, therefore, formed and accumulated in the upper portion of a soil profile, giving a characteristic brown color to soil horizons, the intensity of which decreases with depth. Biogenic calcium and magnesium under such conditions interact with brown humic acids only to a certain extent. Getting into a soil solution, Ca^{2+} and Mg^{2+} partly neutralize fulvic acids which otherwise would “aggressively” tackle the mineral part of the soil, the intensity of podzolic process is weakened and so is soil podzolization.

Brown humic acids, not able to migrate down the profile, interact with sesquioxides, particularly those of iron, forming some stable organic-mineral compounds. Being precipitated mainly in the surface horizons of a soil profile, such complexes favor the formation of brown forest soils with an accumulative type of profile and its more or less clear differentiation into the horizons: $H + H_p + HP + P_h + P$. The distribution of substances over the profile is not distinctly differentiated, and the profile remains of a nondifferentiated, uniformly accumulative type.

Under conditions of lithogenic carbonate accumulation, not all amount of calcium reaches the soil solution. Its considerable part forms a more stable bond with organic acids, compared with the biogenically accumulated calcium, and functions as a kind of a bridge combining humic substances with silicate clay minerals. Simultaneously with the formation of aluminum-iron-humus complexes, a soil profile is being formed well saturated with calcium and other basic cations ($H_k + H_{pk} + P_{hk} + P_k$) without any signs of eluvial differentiation.

On less steep slopes (< 20°), the soluble substances infiltrate and accumulate in deeper soil horizons. Ferrous iron, organic acids, and fine soil particles reach the deeper horizons in alternated anaerobiosis. Losing such materials, the upper portion of soil profile becomes bleached and forms eluvial horizons, whereas in the lower portion of the profile there occurs the accumulation of organic and mineral sediments leached from the overlying horizons.

Illuvial horizons formed in this way are rich in colloids and usually gleyed to the varying extent. The soil profile is clearly divided into eluvial and illuvial portions with reduction (gleying) processes being very operative in podzolization, as these processes increase the solubility of iron, humus, and other substances.

In soils, covering gentle slopes, a continuous leaching of soluble materials and washing of fine particles down the profile causes the development of podzolic processes intensified by reduction processes (gleying-podzolization). In that case, brown forest gleyed podzolic soils are formed which have a clearly differentiated gleyed profile. As the slopes become steeper, gleying podzolization process becomes weaker and less distinctly expressed by morphologic features of a soil profile. On gentle (< 5°) concave and level slopes, very gleyed podzolic soils are formed with the following sequence of horizons: $H_e + H_{pegl} + HP_{igl} + P_{hig}l + P_{gl}$. Moderately and slightly gleyed soil genuses are found on gently steep (10-15°) and moderately steep (15-20°) slopes. Their profiles are composed of the following sequences of horizons: $H_e + H_{pe} + H_{pigl} + P_{higl} + P_{gl}$ and $H_{(e)} + H_{p(e)} + HP_{(i)} + P_{h(i)gl} + P_{gl}$, respectively.

With lithogenic calcium and magnesium accumulation under reducing conditions on slopes less than 20° steep, the infiltrative flow of moisture causes the leaching of these elements downward from the upper portion of a soil profile. Alternative anaerobiosis helps the development of the same eluviation processes as in carbonate-free parent materials, but here they occur in the presence of large amount of calcium. Formed in this way, gleyed and differentiated profile contains calcium and magnesium in some horizons. Such soils, in contrast to gley-podzolic acid brown soils, were named by A.M.Turenko (1989) **gleyed leached**. They form on the same elements of topography but in calcareous or carbonate-containing parent materials. The composition of their profile, depending on the steepness of the slope, may be as follows: $H_e + H_{pegl} + HP_{igl} + P_{hikgl} + P_{kgl}$ (< 5, sometimes 5-10°); $H_e + H_{pe} + H_{pikgl} + P_{hikgl} + P_{kgl}$ (10-15°), and $H_{(e)} + H_{p(e)k} + HP_{ikgl} + P_{h(i)kgl} + P_{kgl}$ (< 20°).

When slopes become less steep, under conditions of periodic anaerobiosis and insufficient drainage, the brown soil formation combines with gleization and leaching of soluble materials (gley-podzolization + gley-leaching). The transfer of substances from the upper horizons is most intensive in soils on the leveled areas of high-level properly drained terraces in Transcarpathia and Subcarpathia. Brown soil formation here is of a asecondary importance. Under such conditions, brown gleyed podzolic soils are formed with the following sequence of horizons in the profile: HE + E_{igl} + I_{gl} + IP_{gl} + P_{igl}.

On low-level terraces with shallow underlying of impervious materials, a continuous influx of clastic materials from mountain slopes and reducing conditions favor the development of alluvial flood-plain processes. While remaining dominating, these processes combine with brow a soil formation. Under such conditions, there is no considerable transfer of soluble materials from the profile owing to poor drainage, and brown-meadow shallow soils are formed. Podzolic soil formation in its typical form is possible only on the slopes with stony colluvial sediments as a parent material.

30.1.2. Brown Meadow Soils

On the floodplains of the river valleys and on the low-level terraces some coarse-grained materials composed of boulders, pebbles, stones, rubble, and sand are deposited during the floods. Silt and clay particles between the stones create a certain layering in the deposition of fine earth. Under grasses and carex-reed associations, in the open areas or under scanty alder forests, shrubs of willow, etc., with a shallow underlying of solid stony rocks, the transfer of fine materials from the shallow (5-50 cm) profile of young alluvial soils is practically negligible. The soil formation is reduced to the accumulation of materials and the formation of grainy structures with aggregates cemented by the brown iron-humate. Iron-humus complexes give to the solum a characteristic brown color.

Brown soil formation in combination with gleization affect the features of alluvial soils, which is evident from the brown color and grainy structure of these soils. Brown meadow soils go through several stages of the soil development from the primitive and underdeveloped to deep full-profiled pedons, which, however, differ from the neighboring soils by the absence of the profile differentiation. Primitive and undeveloped soils formed in alluvial deposits, containing sand, pebbles, and boulders, have a shallow profile (H_d + H + P or H_d + H + $P_{(h)}$ + P), 5 to 20 centimeters deep.

High mountain areas, covered with meadow grasses and shrubs over the upper boundary of forests, are called the mountain meadows or poloninas in Ukrainian. They occupy the tops and slopes of the highest Carpathian mountains (Goverla, Pip Ivan, Petros, Smotrych, Syvula, etc.) in the geomorphologic systems of Chornogory, Svidovetz, Gorgany, and Chivchyny, at the altitudes within 1650-2061 m. The lower boundary of the meadows is rather wavy and, depending on the character of land forms, it may occasionally descend to the altitude of 1200 m.

Mountain meadow vegetation forms several tiers of the canopy cover: cereal species alternating with sedges, lichens, and shrubs. By the character of vegetation, the zone of mountain meadows is divided into three subzones: upper, medium, and lower. The upper subzone occupies the highest level of the Chornogory mountain area (mountains Goverla, Pip Ivan, Petros, Smotrych, Dzembronya, etc) at the altitudes over 1800 m. Some shrub vegetation dominates here, as well as berry species (blackberries and blueberries). There are numerous mosses and lichens. The annual precipitation exceeds 1000 mm. Air temperature is low (8-12°C in July) and the period of plants growth and development does not exceed 90 days. Such conditions are conducive to the formation of peat layers.

The moderate (middle) subzone is situated within the belt of absolute altitudes of 1850-1750 m, with a part of the subzone descending tongue-like to 1400-1000 m, to the upper boundary of forests. Some cereal grass species dominate here, as well as sedges and reeds. There are numerous mosses and lichens. On eroded stony slopes, pine trees may be seen, with curved trunks, 2-3 m in height. They, however, do not form unbroken stands. By climatic parameters, the moderate subzone is warmer than the upper, and to a greater extent sheltered from severe winds. The grass cover is better developed here.

The lower subzone borders directly with the forest zone at altitudes over 1600 m. It forms a narrow broken belt of shrubs and a scanty forest composed mostly of firs. In numerous places, this belt is broken by the tongues of meadow grass lands, which cover sleep slopes and nearly vertical banks of the mountain streams. Brown meadow soils develop here, which by the depth of humus ($H + H_p$) horizons are divided into shallow (20-50 cm), moderately deep (50-70 cm), and deep (> 70 cm). They are gleyed to varying extent ($H_d + H + HP_{gl} + P_{hgl}$).

The Atlass of Soils of the Ukrainian SSR (1979) gives profile description of acid brown forest soils with high OM content on Dumen polonina near Rakhov:

$\mathbf{T}_{\mathbf{d}\ \mathbf{0\text{-}3\ cm}}$: peaty-soddy horizon which consists of semidecomposed plant residues;

$\mathbf{H}_{\mathbf{d}\ \mathbf{4\text{-}12\ cm}}$: soddy-humus horizon; dark grey with brownish tint; light loam; contains rubble; grainy with clods; loose; boundary distinct;

$\mathbf{H}_{\mathbf{13\text{-}40\ cm}}$: humus (accumulation) horizon; brown; heavy loam; more rubble; clods with grains; loose; transition gradual;

HP $_{41\text{-}65\ cm}$: transition (transitive) horizon; light brown; medium loam; cloddy; compact; moderately rich in rubble; transition gradual;

P$_{1h\ 66\text{-}85\ cm}$: fine-grained sandstone eluvium; slightly "dirtied" with SOM; olive-brown; medium loam; cloddy; compact; transition gradual;

P$_{2\ 89\ cm\ and\ deeper}$; sandstone with cracks; platy; olive-gray in color;

Such soils are used for the summer pastures. They need periodic improvement by drilling certain grass mixtures. Liming invigorates the growth of grass. But not overliming. Their pH may be 4.3-4.9. Labile aluminum content (determined by Krupsky's procedure) may be within 363-952 ppm. Hydrolytic acidity is also high (up to 10-12 meq per 100 g of oven-dry soil), and PBS 85-30%. Humus (SOM) content may be as high as 10-12% in the surface (H_d) horizon. SOM is of fulvatic composition (C_{ha} / C_{fa} = 0.53-0.21).

30.1.3. Brown Forest Soils

The classification of brown forest soils is given in Table 104.

Table 104. Classification of Brown Forest Soils

Subtype	Genus	Species
Light brown Dark brown	Modal (acid), carbonate (saturated), gleyed-podzolized, and gleyed-leached	Modal, by the extent of gleying and podzolization, leaching, and erosion

Brown forest soils occur in low, moderate (medium), and high mountain zones. **Dark brown soils** were formed in the subzone of acidulous (fir) forests, at the altitudes of 1250-1650 m. The surface soil contains 8-20% of crude humus. Fulvic acids dominate over humic ones.

Brown forest soils were formed in the subzone of mixed (beech-fir) forests at the altitudes within 600-120 m. SOM content in the surface horizon is within the range of 4-8%, humus variety being moder, which means less crude than in the dark brown soils.

Light brown forest soils are adhering to the mixed forests of the low mountain subzone composed of oak, beech, and hornbeam. Their organic matter is of mull variety which means mild. SOM content seldom exceeds 4%. Iron oxides are dominating in the mineral part of dark brown soils, whereas in light brown ones it is Al_2O_3 that is dominant, which explains the hint of soil color.

The Atlass of Soils of the Ukrainian SSR (1979) gives the following description of the light brown forest soil from Mukachevo district of the Transcarpathian region:

$H_{d\ 0\text{-}6\ cm}$: soddy-humus horizon; darkish grey; heavy loam; grainy with clods; loose; abundant roots and residues; transition clear;

$H_{7\text{-}18\ cm}$; humus horizon; brown when wet, grayish straw-colored-when dry; heavy loam; cloddy with grains; loose; many roots; transition gradual;

$HP_{29\text{-}85\ cm}$: transition horizon; reddish-brown with reddish-black spots; heavy loam; cloddy with nuts; compact; with semiweathered inclusions of solid particles of andesite and basalt; transition gradual;

$P_{85\text{-}150\ cm}$: parent material; deluvium of the cortex of weathering (andesite and basalt); heavy loam; with large clods and nuts; many fragments of solid rock and iron – manganese pointation.

Such soils are very productive for the forests. Where the climatic conditions allow it, the soils may be used for orchards and vineyards. Only small areas are under field crops. A bit of liming can do some good in the latter case. Usually the rates of lime are low and its applications frequent.

30.1.4. Brown Gley-Podzolic Soils

Brown gley-podzolic soils occur on foothill plainlands of the Transcarpathian and Subcarpathian territories, as well as in the hollows between the mountains at the absolute altitudes of 125-400 m.

The soils are formed in non-carbonate loess-like deluvium under mixed (mainly oak-beech) forests on hilly and knolly topography with stagnation and leaching types of soil-water regime. Sufficient drainage on high-level ancient terraces favors an active transfer of substances in a soil profile. But the soil receives a large amount of clastic material from the surrounding mountains containing the same substances as brown forest soils. The conditions of stable anaerobiosis (lack of oxygen) prevail in the soil formation. Gleization (intensive reduction) and material transfer in a soil profile take place under the influence of a brunizemic (brown soil) soil formation.

The morphologic features of a soil profile reflect the character of material migration (percolation, leaching, and other forms of materials transfer) depending on landforms features. The soil diagnostics is performed according to the ratio between the thickness of the upper two soil horizons.

Brown gley-podzolic soils, with profile made up of the horizons: HE + E_{igl} + I_{gl} + I_{pgl} + P_{igl}, are classified into the species of gleyed severely, moderately, and slightly podzolic soils. Different varieties of gleyed severely podzolic soils occupy the level and concave areas of the terraces less than 1° steep. The depth of the HE horizon does not exceeds that of Eigl (HE < E_{igl}). Gleyed moderately podzolic soils have the HE horizon roughly equal to Eigl horizon (HE = E_{igl}), and slightly podzolic soils are characterized by the inequality: HE > E_{igl}. Such soils occur on the terraces with a mainly convex shape of the surface and slopes within 1-3°.

The profile of podzolic brown forest, acid, gleyed from the surface soil, formed in deluvial loam from the Kosiv district of the Ivano-Frankivsk region is described in *the Atlass of Soils of the Ukrainian SSR* (1979).

$H_{e\ (gl)\ 0\text{-}20\ cm:}$ horizon of humus accumulation, eluviated, gleyed, gleyish brown, wet, medium loam, silty, loose, rusty spots and pointations, transition gradual;

HE $_{gln\ 21\text{-}45\ cm;}$ humuso-eluvial horizon; gleyed; dull straw-yellow; medium loam; cloddy with a slight horizontal layering; loose; rusty spots; transition gradual;

$PIGl_{n\ 46\text{-}98\ cm}$; illuvial horizon, very gleyed, yellowish-brown, mottly, heavy loam, prismatic, compact, rustic and iron – manganese pointations, transition gradual;

$P_{migl\ 99\text{-}150\ cm:}$ deluvial loam of a parent material; illuviated; gleyed; straw-yellow with dull gray smears; heavy loam.

Such soils are of low productivity. The practices of their improvement should be directed towards better drainage, liming, and, subsequently, proper fertilizing with both organic manure and mineral fertilizers. The soil pH is within 3.9 and 4.4. Phosphorite meal may prove effective on such soils. Without drainage, it is impossible to obtain the yields of grain over 1 mt per hectare. The available aluminum content is about 145 ppm. The soils may be good for the production of certain medicinal herbs.

30.2. Soils of the Crimean Mountains

30.2.1. Factors and Conditions of Soil Formation

The Crimean mountains occupy the southern part of the peninsula. They are composed of three parallel mountain ridges, each 40-50 kilometers wide and two valleys between the ranges the width of which varies from 2-3 to 15-20 kilometers. The valleys were subject to the impact of erosion processes (Fig. 98).

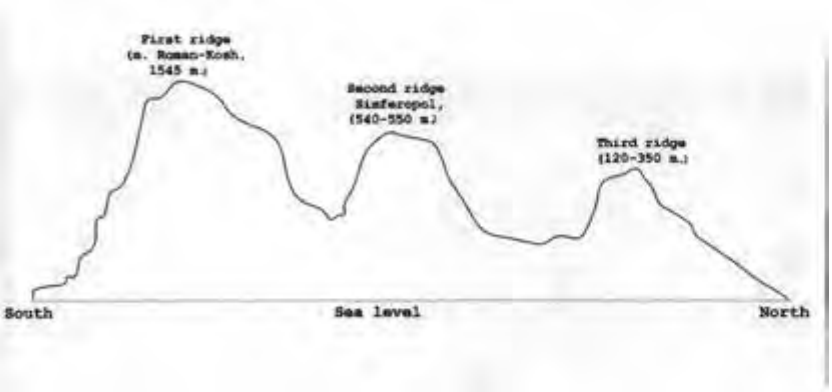


Fig. 98. Schematic Transsection Across the Crimean Mountains (according to M.A. Kochkin)

The height of the ridges lowers from the south to the north. The main (southern) ridge in some places may reach 1540 meters above sea level. The middle ridge may reach 540-550 meters in height and the external (northern) one is only within 120-350 meters high. The northern slopes of the ridges are gentle, gradually descending to the north, whereas the southern slopes are rather steep and, abrupt. Flat mountain tops *(yaylas)* are characterized by a network of broad and deep ravines and gullies. There also are the hills and rocky knolls with crevices and karst funnels, as well as underground wells and streams. Through the crevices and fissures, the snowmelt and rain waters reach the depth of the limestone sediments. The foothill territory is furrowed by river valleys, and other lines of stream pattern, including gullies.

Climate conditions differ in a wide range of characteristics depending on altitudes above the sea level, the slope aspect and steepness, as well as the amount of heat reaching the surface, the temperature of the Black Sea water, and that of the air over the peninsula steppes.

The climate changes very much, depending on the absolute altitude of a location. Every 100 meters upward reduces the air temperature by 0.5-0.6 °C. On the tops of the main mountain ridge at 1250-1450 meters above the sea level, the air temperatures are 7-9 degrees lower compared with foothill areas. The amount of precipitation increases with altitude. Mean annual precipitation is of 460-550 mm at the altitude of 250-300 meters above the sea level, whereas on the tops of the mountains (yaylas) it fluctuates within 900-1250 mm.

The changes of climate with altitude is one of the reasons for the existence of vertical zonality, which becomes evident first of all from the character of vegetation. On the foothills, the areas with needle grass and variegated associations of steppe grasses are bordering with the patches of low-growing forests and shrubs. On the southern and northern slopes of the main mountain ridge, trees dominate over other types of vegetation. Oak forests prevail here, which together with mixed forests dominate at the altitudes of 850-1400 meters above the sea level. Further uphill leads one to the flat tops of the mountains *(yaylas)*, covered by meadow and meadow-steppe grasses occasionally broken by the patches of the Crimean beech and hornbeam.

The Southern coast vegetation is of the Mediterranean character. It is represented by the low-growing mixed forests and shrubs. Occasionally, some oak and juniper forests grow too. Among the conifers, the Sudak and the Crimean pines may be found. Numerous species of ornamental plants make this part of the Crimea a veritable garden of Eden. Vertical changes in climate and vegetation lead to the changes in the character of soil formation. Different soil types are formed in different altitudinal belts, or vertical zones.

Local geology and the character of parent materials also greatly modify the soil cover. The most widely occurring parent materials are eluvium and deluvium of various solid rocks, stony talus, and clastic proluvial deposits. Pebbles, mixed with loams and eluvium-deluvium sediments, are widespread in the river valleys. Loess-like loams and clays are encountered on foothill areas and within the external ridge of the Crimean Mountains.

Factors of soil formation in the mountainous part of the Crimea resulted in the formation of the following vertical zones: 1) foothill Steppe zone of southern chernozems and meadow chernozemic soils; 2) foothill forest-Steppe zone of soddy carbonate (rendzinas) and grey forest mountain soils; 3) mountain forest zone of brown forest soils; 4) mountain meadow zone on *yaylas* with mountain-meadow chernozem-like soils, and 5) southern slopes of the main ridge covered by cinnamonic soils. Brown forest soils in the Carpathians are known as Cambisols Eutric in the FAO soil nomenclature. Cinnamonic mountain soils formed mainly in hard rock eluvium-deluvium are known as Cambisols Calcaric in the FAO soil nomenclature.

Eroded soils frequently occur on mountain slopes, as well as talus slopes rich in stones and rubble.

30.2.2. Soddy-Carbonate Mountain Forest-Steppe Soils (Rendzinas Cambic)

Soddy-carbonate mountain Forest-steppe soils are widespread in the Forest-Steppe zone of the foothill area. Considerable areas of such soils locate on the slopes of internal and, to a lesser extent, of the main and external mountain ridges up to 400-450 meters above the sea level. These low-mountain soils were formed under the steppe grasses and the associations of trees and shrubs in eluvium-deluvium of solid chalk-marls and limestones of neogenic and paleogenic age. The soils have shallow and stony profiles, 40-60 cm deep. The horizon of OM accumulation (H_k) contains 2-6% of SOM and is of cloddy or cloddy with crumbs aggregation. The carbonate content may reach 30%, and the soil reaction is slightly alkaline (pH $_{H2O}$ 7.3-8.0).

Full-profiled and eroded (or xeromorphic) mapping units were developed in the eluvium of limestones, whereas the profiles covered by eroded materials – in deluvium of the same solid rocks. A brief soil profile description may be as follows: horizon of OM accumulation (H_k): 5-25 cm deep, grey, skeletal, loose, loam, cloddy, transition gradual; upper transition horizon (H_{pk}): 15-20 cm thick; light grey, skeletal, compact, loam, cloddy, transition clear; lower transition horizon (P_{hk}): 20-25 cm thick, light-colored, very skeletal; descend into slightly weathered limestone. In profiles less than 50 cm deep only one transition horizon is identified. Soils covered by eroded materials may be 90-120 cm deep.

By the depth of loose weathered layer (H_k + H_{pk}), the soddy-carbonate mountain forest-steppe soils are classified into undeveloped (< 25 cm), short profiled (25-45 cm), and modal (> 45 cm). By OM content, they are divided into those with slight and those with low OM content (< 3% and 3-6%, respectively). There also are soil species identified by the extent of soil erosion. With the greater erosion the soils become more skeletal, but by the intensity of this structure property the soils are also divided into species, as slightly, moderately, and very stony.

Very stony species may contain over 30% of skeleton particles. Such percentage is determined relative to the area of the soil horizon, and not to the soil weight. Soils, covered by the eroded material, are found on the lower portions of slopes. Their profile may be composed of the following horizons: $H_{k,\ 50\text{-}60\ cm}$ thick; + $H_{pk\ 25\text{-}40}$ + $P_{hk\ 35\text{-}40}$ + P_k.

30.2.3. Grey Mountain Forest-Steppe Soils

Grey mountain Forest-Steppe soils of the vertical Forest-Steppe zone occupying a band-like area from 350 to 600-900 meters above the sea level mainly on the slopes of southern and south-western aspects and the tops of mountain ranges. **Vegetation** species and associations favoring the formation of such soils are represented by trees, shrubs, and grasses. **Dominant parent materials** are represented by eluvium and deluvium of paleogenic limestones, shales, and slates and their weathering and transportation products. A characteristic feature of soils is their grey (or dark grey) color in the surface horizon of OM accumulation. This horizon is of a friable nutty and prismatic structure and of illuvial character, which means enriched with clay and colloidal matter, which can be detected by the particle size analysis: the amount of clay at the depth of 45-65 cm is by 3-10% larger than in the surface horizon (0-20 cm).

Grey forest soils formed in limy materials show effervescence from the depth of 35-55 cm. The depth of soil profile typically is within 60-100 cm. By texture, the soils are clays or clay loams. They contain from 3 to 5 % of SOM. The soil reaction is slightly acid or close to neutral (pH $_{H2O}$ 6-7).

A typical soil profile may be characterized by the following morphologic features: H, 25-30 cm thick grey or dark grey; compact; grainy; gradually passes into the upper transition horizon (H_p), 20-25 cm thick, grey with brownish tint, compact; clay; cloddy with nuts; inclusions of rubble; effervesces in the lower portion; transition gradual; lower transition horizon ($P_{h/k}$), 25-30 cm thick; light grey; compact; clay; stony; occasional carbonate micellia; transition to parent material distinct. The soil species are identified by the extent of soil destruction by erosion. Washed-on species are formed on lowland areas and lower portions of slopes in deluvial deposits. Their profile may be 100-110 cm deep.

30.2.4. Meadow Mountain Chernozem-like Soils

Meadow mountain chernozem-like soils occur on flat mountain tops *(yaylas)* free from forests and partly on the slopes of the main mountain ridge, 600 meters above the sea level. The soils were formed in increased humidity and decreased amount of heat, under meadow and meadow-steppe grasses, in the products of weathering of limestones, conglomerates, and sandstones or, occasionally in the consolidated clay of karst funnels. On elevated level yayla areas, the soil profile is 30-55 cm thick, the surface horizon of SOM accumulation being of 8-20 cm. By texture, the soils are mostly clay loams. They show effervescence with dilute HCl from the surface or the depth of 10-15 cm. On areas, sloping to the north, and on low patches, the soils are leached from carbonates. SOM content is within 2-8%. Skeletal particles occupy 30-40% of the area of soil horizons. Their content increases downward. The soil adsorbing complex is well saturated with calcium, the soil reaction being close to neutral.

The soils formed in the eluvium of limestone are calcareous, whereas those in conglomerate and sandstone eluvium are noncalcareous, and those formed in clayey deluvium of limestones, are mostly leached from carbonates. Of calcareous soils, the most widespread ones are shallow-profiled. Their humus-accumulation horizon (H_k), 8-10 cm deep, is dark grey, stony, compact, grainy or cloddy in structure. The upper transition horizon (H_{pk}) is 15-20 cm thick, grey, stony, very compact, and cloddy. The lower transition horizon (P_{hk}) is 18-25 cm thick, light grey and very stony. In profiles as shallow as 30-35 cm, only one transition horizon is identified.

Leached genuses of meadow mountain chernozem-like soils occupy only small areas. Their profile is indistinctly differentiated, 120-140 cm thick, with SOM content of 8-10% and the soil reaction close to neutrality.

H - horizon is 20-40 cm thick, dark grey, loose and friable, with cloddy and grainy aggregations, H_p - horizon is 40-45 cm thick, grey, compact, and cloddy; P_h – horizon is 40-55 cm thick, light grey, compact, and nutty in aggregation.

By the depth of the profile of SOM accumulation (H + H_p), the soils are classified into undeveloped (< 25 cm), moderately deep (65-85 cm), deep (85-120 cm), and extra deep (> 120 cm). By SOM content, they may be slightly (< 3%), lowly (3-6%), and moderately (> 6%) humusified.

30.2.5. Brown Forest Soils

Brown Forest Soils are widespread in the forest mountain zone of the main and internal mountain ridges at the altitude above 300 meters on the northern-aspect slopes and over 400-600 meters on the southern-aspect slopes and the upper boundary within 900-1200 meters above the sea level. The annual precipitation fluctuates within 550-1,150 mm.

Brown forest soils of the Crimea were formed mainly in the eluvium and deluvium of limestones, shales, slates, sandstones, conglomerates, and, occasionally, hard crystalline rocks. The intensity of soil formation depends on parent materials and forest types. It reaches its acme in noncarbonated parent materials under beech, hornbeam, and beech- hornbeam forests, whereas pine forests are less conducive to its development.

Sufficient humidity, together with prolonged warm period, causes a rapid disintegration and decomposition of primary minerals, as well as the synthesis of secondary ones under periodic reduction conditions. The accumulation of secondary minerals in a soil profile is accompanied by transition of soluble materials via percolation and the lateral flow of moisture in the soil.

Soluble products of soil formation, including carbonates and clay particles, are thus washed off the upper portion of the profile. Rapid mineralization of organic residues and humus substances are in the way of soil humus accumulation. An increased amount (4-6%) of SOM may be determined only in the upper part of the surface horizon, right under the forest litter. SOM content rapidly decreases with depth. Soil humus is of fulvatic type all over the profile.

Clay loam soil varieties with stones and rubble in the lower portion of soil profile are dominant. CEC is within 25-40 meq. per 100 g of oven-dry soil. Calcium is a dominant exchange cation. Soil reaction may be slightly acid in soils formed in noncarbonatic parent materials (pH $_{H_2O}$ 5.2-6.8) and neutral to weakly alkaline (pH $_{H_2O}$ 6.8-8.0) in soils formed in carbonate parent materials.

A slightly podzolized brown forest soil in the eluvium of shales in Alushta district at the altitude of 752 meters above sea level, described in *the Atlass of Soils of the Ukrainian SSR* (1979), has the following morphologic features:

$H_{0\ 0\text{-}2\ cm}$: forest litter;

$H_{d(e)\ 3\text{-}12\ cm}$: soddy horizon; darkish-brown; clay loam; slightly stony (skeletal); cloddy with grains; loose; porous; with cracks; abundant roots; transition clear;

$H_{(e)\ 13\text{-}25\ cm}$: humus horizon; slightly eluviated; light brown; light clay; slightly skeletal; cloddy; loose; porous; with cracks; many roots; transition gradual;

$H_{pi(gl)\ 26\text{-}40\ cm}$: noticeably illuviated; yellowish brown; slight olive shade; light clay; slightly skeletal; nutty; with cracks; rusty spots on the faces of soil clods; roots; transition gradual, with tongue-shaped line.

$PI_{hgl\ 41\text{-}100\ cm}$: brownish dull-gray; numerous rusty spots; light clay; nutty with clods; compact; rich in skeletal particles of shade which give a darker color to the horizon; with thin cracks; scanty roots; transition gradual;

$P_{(gl)k\ 101\text{-}170\ cm}$: eluvium of shale of a nonuniform composition, effervesces from the depth of 125 cm.

Such soils are mainly under beech forests, the productivity of which may be within 500-600 cubic meters of timber (M.A.Kochkin, 1967). Practices of forest protection and soil conservation are a must here. Forest cover usually keeps erosion to very low levels if trees are harvested by the selection method. The logs removal may not only be by tractor but by cable or helicopter. Forest fertilization here also deserves attention.

30.2.6. Cinnamonic Soils of Dry Forests and Shrubberies

The entire area of cinnamonic soils of dry forests and shrubberies is only about 42 thousand hectares of which only 7.2 thousand hectares are a plowland. The soils occur on the southern slopes of the main ridge of the Crimean Mountains at the altitude of 350-500 meters above sea level. These areas are the subtropics of Ukraine. The cortex of weathering is allitic here and the secondary minerals may dominate over the primary ones. The soils are mostly brownish-red. The climate is of the Mediterranean type with droughty summer and mild wet winter. The annual precipitation in this part of the Crimea is within a wide range of 300-565 mm. The subtropical vegetation is similar to that of Southern Europe: evergreen shrubs and trees, such as myrtle, oleander, laurel and evergreen oak. The forests are thin but biodiverse with maple, beech, chestnut, and walnut.

Limestones, shales, and conglomerates (their eluviums) are the dominant parent materials. The soil profile described in *the Atlass of Soils of the Ukrainian SSR* (1979) near Alushta at the altitude of 300 meters above the see level has the following morphologic features:

$\mathbf{H_{0\ 0\text{-}2\ cm:}}$ forest and grass litter; fresh and semidecomposed residues;

$\mathbf{H_{(e)d\ 3\text{-}14\ cm:}}$ soddy horizon; eliviated; brown-grey; heavy loam; very skeletal; grainy; compact; porous; abundant roots; large rubble of shale; transition clear;

$\mathbf{H_{(i)\ 15\text{-}30\ cm:}}$ horizon of humus accumulation; slightly illuviated; cinnamon in color; light clay; very skeletal; grainy with clods; compact; abundant roots; transition gradual;

$\mathbf{H_{p\ 31\text{-}55\ cm:}}$ upper transition horizon; cinnamon with brownish tint; light clay; very skeletal; cloddy; very compact; some violet and dull grey spots of local gleying; transition gradual;

$\mathbf{P_{hk\ 56\text{-}80\ cm:}}$ lower transition horizon; cinnamonic-brown; light clay; large clods with nuts; very compact; over 50% of skeleton; effervesces with 10% HCl; carbonate micellium; transition clear;

$\mathbf{P_{k\ 80\text{-}170\ cm:}}$ eluvium of shale.

Red-cinnamonic soils were described and photographed (Atlas, 1979) in Yalta district (Nikitsky Botanical Garden). The profile consists of the horizons: $H_0 + H_{(k)} + H_{p(k)} + P_{hk} + P_k$. The soil was formed in limestone deluvium.

Cinnamonic and red-cinnamonic soils are good for vineyards, tobacco plantations and other subtropical crop production. Forest protection and soil erosion control are the basic prerequisites of soil management.

Questions & Assignments

1. What is a vertical zonality? What is the inversion of soil zones?
2. Describe the general features of Soil formation in the Carpathians.
3. What are the essential features of brown forest soil formation?
4. Speak on the use of brown meadow soils.
5. Speak on the classification of brown forest soils.
6. Describe specific features of soil formation in the Crimean mountain region.
7. Describe the profile of a meadow mountain chernozemic soil.
8. What is the difference between the brown forest soils of the Crimea and the Carpathians?
9. Speak on the practices of soil conservation in the mountains.
10. Describe Cinnamonic soils of dry forests and shrubberies.

Chapter 31

Soil Erosion and its Control

Soil erosion is a natural or anthropogenic process opposing the soil formation. Erosion occurs in many forms as a result of natural and anthropogenic causes. Erosion in natural ecosystems caused by natural agents is called **normal or geologic.** The normal rate of erosion under natural vegetation is in approximate equilibrium with the rate of the soil formation. Precise data on rates on geologic erosion and the soil formation are difficult to obtain but they are thought to average about 1 mt/ha annually (F.R.Troech, I.A.Hobbs, and R.L.Donahue, 1991). Tilling cropland, grazing pastures or rangeland, or cutting trees nearly always increases the rate of soil erosion. Loss of soil cover reduces protection and may accelerate soil loss by a factor of 10 to 100 times. The formation of a new soil cannot keep pace with greatly accelerated erosion rates.

A noted Ukrainian soil conservationist M.K. Shikula proposed to estimate the rate of soil erosion by the following scale (1973):

Erosion Rate, mt/ha• yr:	Its Estimation:
0-3	Soil erosion absent
3-6	Weak soil erosion
6-12	Moderate soil erosion
12-24	Severe soil erosion
24-60	Very severe soil erosion
> 60	Catastrophic soil erosion

Accelerated erosion is thought by some Ukrainian experts as the only form of erosion worth investigation and control (S.Yu.Bulygin) or at least the only form of erosion worthy of prevention. Accelerated erosion reduces the amount of plant growth a soil is able to support. According to F.R.Troech, et al (1991), the productive potential is reduced even if the actual production is maintained or increased by the use of fertilizers and other management techniques. A shallower soil, with its reduced capacity for storing water and plant nutrients and its generally poorer structure and aeration, cannot match the productive potential of the uneroded soil. There is a need for erosion control practices that will reduce the erosion rate and maintain the soil productivity.

31.1. Water Erosion

Any erosion (caused by either water or wind) is a three-step process: **detachment**, followed by **transport** and **deposition**. The energy of water erosion is derived from falling rain and the movement of runoff water caused by rain or show melting. Irrigation may also cause water erosion.

Several types of water erosion have been identified: raindrop (splash), sheets, rill, and gully or channel (H.D.Foth, 1990). In addition, undercutting of stream banks by water causes large masses of soil to fall into the stream and be carried away. Very wet soil masses on steep slopes are likely to slide slowly downhill (landslides). In the Ukrainian classification there are two basic forms of water erosion – surface and linear. Gullies belong to linear erosion as well as large rills which cannot be erased by ordinary tillage. Sheet and not too deep rill erosion belong to the surface type. The removal of soil material from the sides of running streams is called **stream bank erosion.**

Sheet erosion is the removal of thin layers of soil over the whole soil surface. Raindrop splash and surface flow cause sheet erosion. Splash provides most of the detaching energy. Flow provides most of the transporting capacity. Sheet erosion may be difficult to see. The first sign is when the subsoil color begins to show, as the cultivation mixes surface soil and subsoil. It is most apparent on upper portions of convex slopes.

Rills are channels small enough to be obliterated by normal tillage operations. Rill erosion occurs when runoff water concentrates in streamlets as it passes downhill.

Gully erosion is the formation of erosion channels too large to be erased by ordinary tillage. The channels in deep loess soils are U-shaped with almost vertical walls. Broad V-shaped channels often develop where friable surface soils overlie cohesive, tight, nonerodible subsoils. **Gullies** are active when their walls are free of vegetation, and **inactive** when they are stabilized by vegetation. Gullies are also classified as small, medium, and large according to their depth, with medium-sized gullies 1 to 5 m deep.

31.1.1. Water Erosion Factors and Conditions

M.K.Shikula, et al. (2001, 2004) identified the following natural agents of soil erosion: 1) landforms. 2) shape of slopes; 3) aspects of slopes; 4) steepness of slopes; 5) soil erodibility; 6) size of natural watersheds; 7) intensity and length of rain; 8) amount of water in the snow, and 9) the speed of snowmelting in spring. But these natural agents may be complimented by especially dangerous anthropogenic agents including 1) great length of the field along the slope, 2) great size of artificial watersheds, 3) incorrect position of the elements (borderlines) of territory organization, 4) wrong direction of soil tillage; 5) use of moldboard plows, 6) overloading of crop rotations with row crops, 7) nonuniform snow trapping, etc.

Potential erosional damage (PED) may be expressed by the equation:

$$PED = f(Cl, T, G, S, V, LU) \quad (77)$$

where f is the sign of a function; Cl = climate; T = topography; G = geologic conditions; S = soil conditions; V = vegetation, and LU = character of land use.

H.D.Foth (1990), speaking on predicting water erosion rates on agricultural land, states that the rate of soil erosion on agricultural land is affected by rainfall characteristics, soil erodibility, slope characteristics, and vegetative cover, as well as management practices. Erosion rates in the USA were and still continue to be predicted by using the Universal Soil-Loss Equation (USLE) developed by Wischmeier and Smith. The USLE equation is as follows:

$$A = RKLSCP \quad (78)$$

where A is the computed soil loss per unit area as tons per acre, R is the rainfall factor, K is the soil-erodibility factor, L is the slope-length factor, S is the slope-gradient factor, C is the cropping-management factor, and P is the erosion-control practice factor. The equation is designed to predict water erosion rates on agricultural land surfaces, exclusive of erosion resulting from the formation of large gullies.

The rainfall factor (R) is a measure of the erosive force of specific rainfall. Available energy is related to both the quantity and intensity of rainfall.

V.V.Slastikhin (1964) proposed to calculate the erosive force of rain (Δ) by the formula:

$$\Delta = i\sqrt{t} \quad (79)$$

where i is the intensity of rain in mm•min^{-1}, and t is the time length of that rain in minutes. By the erosive force (Δ), the rains are classified as shown in Table 105:

Table 105. Classification of the erosive force

Type of rain	Erosive force	Consequence of rain
Drizzle	Up to 1	No runoff, only some very slight raindrop erosion is possible
Ordinary	1-3	Slight runoff, slight soil detachment
Slight shower	3-5	Noticeable runoff and noticeable soil loss on the slopes
Moderate shower	5-7	Numerous streamlets on the slopes and severe sheet and rill erosion
Strong shower	7-9	Inundation of floodplains, very severe rill and gully erosion
Very strong shower	9-12	Floods in the basins of small rivers, very severe rill and gully erosion, active landslides

H.D. Foth in his *Fundamentals of Soil Science* notes that a 5-centimeter rain falling at 32 kilometers per hour (nearly 9 m•sec^{-1}) would have a kinetic energy sufficient to raise an 18-centimeter-thick furrow slice to a height of 1 meter. Anything that protects the soil from raindrop impact, such as plant cover, protects the soil from erosion. **The rainfall factor (R)** is the product of the total kinetic energy of the storms times the maximum 30-minute intensity of fall and modified by any influence of snowmelt.

The soil erodibility factor (K) depends on 1) those factors that affect infiltration rate, permeability, and total water retention capacity and 2) those factors that allow soil to resist dispersion, splashing, abrasion, and the transporting forces of rainfall and runoff.

LS = The Slope Length and Slope Gradient Factors. The slope length is defined as the distance from the point of origin of overland water flow either to a point where the slope decreases to the extent that deposition occurs, or a point where water runoff enters a well-defined channel. The runoff from the upper part of a slope contributes to the total amount of the runoff that occurs on the lower part of the slope. This increases the quantity of water running over the lower part of the slope thus increasing erosion more on the lower part of the slope than on the upper part. Studies have shown that erosion via water increases as the 0.5 power of the slope length ($L^{0.5}$). This results in about 1.3 times greater average soil loss per acre for every doubling of slope length.

As the slope gradient (percentage of slope) (S) increases, the velocity of the runoff water increases, which in turn increases the water's erosive power. The doubling of runoff water velocity increases erosive power four times, and causes a 32-time increase in the amount of material of a given particle size that can be carried in the runoff water.

The cropping-management factor (C). A vegetative cover can absorb the kinetic energy of falling raindrops and diffuse the rain's erosive potential. The vegetation retains a significant amount of rain and shows the flow of runoff water. The presence or absence of a vegetative cover determines whether erosion will be a serious problem. O.P.Verveyko (1971) calculated the coefficient of erosional hazard connected with various field crops and summer fallow. M.N.Zaslavsky subtracted them from the unity and in this simple way obtained the coefficients of antierosional efficiency of the same crops (Table 106).

The benefit of a rotation is that a year of high erosion, during the corn production year, is averaged with years of lower erosion, when corn is grown one year followed by wheat, which is then followed by a hay or meadow crop. However, when corn is grown no-till, there may be little soil erosion because of the protection of the soil by previous crop residues.

In general, wherever the sloping land is cultivated and exposed to erosive rains, the protection offered by sod or closely grown crops in the system needs to be supported by practices that will slow runoff water, thus reducing the amount of soil carried. Limited field studies in the USA (H.D.Foth, 1990) have shown that contour farming alone is effective in controlling erosion during rainstorms of low or moderate intensity, but it provides little protection against occasional severe storms that cause breakovers by runoff waters of the contoured ridges or rows.

Table 106. Coefficients of Erosion Hazard and Antierosional Efficiency of Field Crops and Summer Fallow

Field of a Crop Rotation	Coefficient of erosion hazard	Coefficient of antierosional efficiency
Summer fallow	1	0
Corn for green fodder	0.75	0.25
Corn with soybean or peas	0.50	0.50
Corn for silage	0.60	0.40
Corn for grain	0.85	0.15
Sugar beets	0.85	0.15
Potatoes	0.85	0.15
Sunflowers	0.75	0.25
Spring small grain crops	0.50	0.50
Winter small grain crops	0.30	0.70
Perennial grasses: first year	0.08	0.92
second year	0.03	0.97
third year	0.01	0.99

Strip cropping together with contour tillage, provides more protection. In cases where both strip cropping and contour tillage are used, the P values in the USLE are divided by 2. Terraces are an effective way to reduce slope length, which becomes the distance between the terraces.

31.1.2. Water Erosion Damage and Control

There are many ways in which water erosion causes damage. The soil is lost, plant nutrients are removed, the texture changes, the structure deteriorates, the productive capacity is reduced, fields are dissected, and sediments pollute streams and lakes and pile up on bottomlands, in stream channels, and in lakes and reservoirs (F.R.Troeh, et.al, 1991). Not so recent estimation (Colacicco et al., 1989) of the annual damage caused by water erosion in the USA are: on-site damages – 500 million to $1.2 billion; off-site damages – $ 3.4 billion to $ 13 billion. Here in Ukraine each hectare of a plowland loses 15-18 mt/yr of surface soil, whereas tolerable soil loss should not exceed 2-3 mt/ha×yr on soddy-podzolic and 5 mt/ha×yr on chernozemic soils. Erosion also damages buildings, roads, bridges, and other engineering structures. Soil sediment is the greatest single pollutant of surface water. Troeh et al., (1991) formulate the principles of water erosion control in a very simple way: 1) reducing raindrop impact on the soil; 2) reducing runoff volume and velocity, and 3) increasing the soil's resistance to erosion.

Many different practices have been developed to reduce water erosion. Not all practices are applicable in all regions. Soil conservation begins with a proper soil survey and the land use planning. The land should be used according to its capability. Plowland should not extend itself to slopes over 5° (9%) and row crops should not be cultivated on slopes over 3° (5%). Slopes over 9% should be used for grasslands, pastures and afforestation. Conservation tillage should become a must on arable lands and include a proper residue cover management up to no-till where necessary. The proper land use planning should be based on contour farming. Multiple cropping and strip cropping should compliment contour farming. The conservation structures, including terraces and diversions, grassed waterways, gully control structures, earthen dams, etc. should be employed where necessary. Wind breaks and shelterbelts as well as vegetating mining and conservation sites should be employed systematically. That will help to compliment the soil conservation with the soil and water pollution control. Troeh et al., (1991) call their reader's attention to the fact that the historic soil erosion rates have changed as the land use has changed. Much of the land in the eastern United States, which contributed much sediment to streams and rivers a hundred or more years ago, has been removed from cropping. The lower erosion rates resulted from the abandonment of land for cropping and returning it to forest. But in the past few decades, much of this land has experienced a high erosion rate because of urbanization, construction of shopping centers, roads, and housing areas. Such an experience should not go unnoticed by the Ukrainian land planners as the area of the urban lands here is rapidly increasing.

31.2. Wind Erosion

According to H.D.Foth (1990), wind erosion is indirectly related to water conservation in that a lack of water leaves the land barren and exposed to the wind. Wind erosion is greatest in semiarid and arid regions. It is so when the form of wind erosion is a dust storm. But in Ukraine, even in the semihumid areas of the Forest zone, a slight form of wind erosion occurs locally known as *"pozemka"* on very light textured soils used as cropland. Another slight form of wind erosion in Ukraine occurs in the valleys of the rivers with drained and "overdried" peat soils. This form of erosion – a small dust storm along the river valley – takes place mainly in the Forest zone (Polissya).

Soil particles move in three ways during wind erosion: **saltation, suspension, and surface creep.** During saltation, fine soil particles (0.1 to 0.5 mm in diameter) are rolled over the surface by the direct wind pressure, then suddenly jump up almost vertically over a short distance to a height of 20 to 30 centimeters. Once in the air, the particles gain in velocity and then descend in an almost straight line, not varying more than 5 to 12 degrees from the contour line. The horizontal distance traveled by a particle is four to five times the height of its jump. On striking the surface, the particles may rebound into the air or knock other particles into the air before coming to rest. About 90 percent of the total soil movement via wind takes place below a height of 30 centimeters. Probably 50 percent or more of movement occurs between 0 and 5 centimeters above the soil surface (H.D.Foth, 1990). Very fine dust particles are protected from the wind action, because they are too small to protrude above a minute viscous layer of air that clings to the soil surface. As a result, a soil composed of extremely fine particles is resistant to wind erosion. These dust-sized particles are thrown into the air chiefly by the impact of particles moving in saltation. They may be carried very high and over long distances via suspension. Relatively large sand particles (between 0.5 and 1.0 mm in diameter) are too heavy to be lifted by wind action, but they are rolled or pushed along the surface by the impact of particles during saltation. Their movement is by surface creep.

From these facts, it is evident that wind erosion goes on due principally to the effect of wind on particles of a size suitable to move in saltation. Accordingly, wind erosion can be controlled if: 1) soil particles can be built up into peds or granules too large in size to move by saltation: 2) the wind velocity near the soil surface is reduced by ridging the land, by a vegetative cover, or even by developing a cloddy surface; and 3) strips of stubble or other vegetative cover are sufficient to catch and hold particles moving in saltation (Troech, et al., 1991).

31.2.1. Wind Erosion Affecting Factors

The wind erosion prediction equation used in the USA for a long time had been proposed by N.P. Woodruff and F.H. Siddoway in 1965. It is still being refined and is in the form of a function:

$$E = f(I, K, C, L, V) \tag{80}$$

where: E equals the soil loss in tons per acre, I equals the soil erodibility; K equals the soil roughness; C equals the climatic factor; L equals the field length, and V equals the quantity of vegetative cover. But the use of this equation is more complex than that of USLE and the details are given in *the National Agronomy Manual* of the SCS, 1988 or in Troeh et al, 1991.

The soil erodibility by wind is related primarily to the soil texture and its structure. As the clay content of soils increases, the increased aggregation creates clods or peds too large to be transported. If a surface layer of a soil (say, two inches deep) contains not less than 60 percent by volume of soil particles and aggregates over 1 mm in diameter, the soil surface is resistant to wind erosion.

Rough soil surfaces reduce wind erosion by reducing the surface wind velocity and by trapping soil particles. The surface roughness can be increased with tillage operations. For maximum effectiveness, the tillage and rows are positioned at right angles to the dominant wind direction.

Climate is a factor in wind erosion via its effect on wind frequency and the velocity and wetness of soil during high-wind periods. Wind erosion and blowing dust are persistent features of deserts in which plants are widely spaced on and dry soil is exposed most of the time. Wind erosion preferentially removes the finer particles, leaving the larger particles behind. In this way a **desert pavement** is created, consisting of gravel particles that are too large and heavy to be blown away. Deserts are an important source of loess, which is composed mainly of silt-sized particles. Climate effects the amount size of a vegetative cover. During severe dust storms (Dust Bowl) of the 1930s in the USA and in the late fifties – early sixties on the virgin lands of Northern Kazakhstan, both yields of wheat and wind erosion were related to rainfall. The lowest yields and the highest soil erosion rates occurred during the years of least rainfall.

Wind erosion increases from zero, at the edge of an open field, to a maximum, with the increasing distanc**e** of soil exposed to the wind (length of the field factor). As particles bounce and skip, they create an avalanche effect on the soil movement. E.I.Shyiaty (1986), from the former All-Union Institute of Grain Farming, established a relationship between the rate of deflation and the parameters of the soil surface:

$$Q = 10^{a-bk-cs} \tag{81}$$

where Q is the rate of soil detachment, g·m^{-2}; K equals the content (percentage) of soil particles over 1 mm in diameter in the surface 2-inch layer of the soil; S equals the amount of items of plant residues per square meter, and a, b, and c are empirical coefficients dependent on the soil texture.

31.2.2. Wind Erosion Control

Contour farming is effective for wind erosion control. Wind-breaks of trees and alternate strips of crops can be used to reduce the effective field length exposed to the wind. Wind erosion is inversely related to the degree of a vegetative cover. Growing crops and the residues of previous crops are effective in controlling wind erosion.

The stubble mulch tillage – the tillage that leaves plant residues on the soil surface complimented by shelterbelts is popular in wheat producing regions with a moisture-deficient climate.

Deep plowing has been used to control wind erosion where sandy surface soils are underlain by spodic horizons that containing 20-to 40 percent clay. This kind of plowing was used in the USA (Kanzas and Texas) and increased the clay content of the surface soil by 5 to 12 percent in some cases (Foth, 1990). Since 27 percent clay in the surface was required to halt soil blowing, deep plowing itself resulted in only partial and temporary control of wind erosion. When deep plowing was not accompanied by other soil erosion control practices, wind erosion removed clay from the surface soil, and the effects of deep plowing were temporary.

Organic soil areas of appreciable size are frequently protected from wind erosion by planting trees in rows-windbreaks. So it is done in the USA. But such practice must be limited, however, because of the large amount of soil taken out of crop production. Shrubs are also used for the purpose. For example, spirea is popular as a wind break. Trees and shrubs also provide cover for wild life. The lightness of dry organic soils contributes to the wind erosion hazard. The moist organic soil is heavier and particles tend to adhere more. That is why the depth of watertable on dried (drained) peats should be maintained within 0.9-1.5 meters. Use of an overhead sprinkler irrigation system is helpful during windy periods, before the crop cover is sufficient to provide protection from the wind. Rolling the muck soil with a heavy roller induces moisture to rise more rapidly by capillarity, thus dampening the soil surface. In conclusion, it would be reasonable to state that wind erosion control is based on one or more of the following:

1) trapping soil particles with rough surface and the use of crop residues and strip cropping.

2) deep plowing to increase the clay content of the surface soil and the simultaneous use of other erosion control practices.

3) protecting the soil surface with a vegetative cover. Some farmers are wise enough to start moldboard plowing with the beginning of a wind storm. This practice increases the surface soil moisture and the humidity of the surface air, which rapidly increase the critical wind velocity, and a wind storm does not start on a plowed area. In Ukraine the employment of windbreaks and shelterbelts to protect fields is ubiquitous and systematic.

Questions & Assignments

1. What is a geologic soil erosion?
2. How do you propose to estimate the intensity of water erosion?
3. What is a tolerable soil loss? What factors does it depend on?
4. Speak on the types of water erosion.
5. What factors of water erosion were identified by M.K.Shikula?
6. What is your opinion of the USLE?
7. What practices of soil management can reduce soil erodibility factor?
8. Characterize cropping – management factor. How can it be utilized in planning crop rotations?
9. Name the practices of water erosion control.
10. Name the factors affecting wind erosion.
11. Name the basic practices of wind erosion control.
12. Comment on the extent of erosion damage in Ukraine.

LITERATURE

1. Атлас почв Украинской ССР. Под ред. Н.К.Крупского и Н.И.Полупана. К.: Урожай. – 1979. – 160 с.
2. Булигін С.Ю. Оцінка і прогноз якості земель: Навч. посібник / С.Ю. Булигін, А.В. Барвінський, А.О. Ачасова, А.Б. Ачасов / Харк. нац. аграр. ун-т. – Х., 2008 . – 237 с.
3. Булыгин С.Ю. Почвоводоохранная оптимизация агроландшафтов: Монография / С.Ю. Булыгин, В.А. Белолипский. – К.: Аграр. наука, 2012. – 352 с.
4. Величко В.А. Екологія родючості ґрунтів / В.А. Величко. – К.: Аграрна наука, 2010. – 271 с.
5. Воронин А.Д. Основы физики почв / А.Д. Воронин. – М.: Изд. МГУ. – 1986. – 244 с.
6. Гнатенко О.Ф. Ґрунтознавство. Лабораторний практикум / О.Ф. Гнатенко, Л.Р. Петренко, М.В. Капштик та ін. – К.: РВЦ НАУ. – 2000. – 170 с.
7. Гнатенко О.Ф. Практикум з ґрунтознавства / О.Ф. Гнатенко, Л.Р. Петренко, М.В. Капштик та ін. – К.: РВЦ НАУ. – 2002. – 230 с.
8. Гнатенко О.Ф. Ґрунтознавство з основами геології / О.Ф. Гнатенко, М.В. Капштик, Л.Р. Петренко, С.В. Вітвицький. К.: Оранта. – 2005. – 648 с.
9. Кауричев И.С. Почвоведение / И.С. Кауричев, Н.П. Панов, Н.Н. Розов и др. Под. Ред. Кауричева. 4-е изд. – М.: Агропромиздат. – 1989. – 719 с.
10. Медведев В.В. Физическая деградация черноземов. Диагностика. Причины. Следствия. Предупреждение / В.В. Медведев. – Харьков: Изд-во «Городская типография», 2013. – 324 с.
11. Національна доповідь про стан родючості ґрунтів України / За ред. С.А. Балюка, В.В. Медведєва і А.Д. Балаєва. – К.: НААНУ. – 2010. – 153 с.
12. Орлов Д.С. Химия почв / Д.С. Орлов. – М.: Изд-во МГУ. – 1985. – 376 с.
13. Петренко Л.Р. Зміна біологічних властивостей ґрунтів під впливом обробітку ґрунту без обертання скиби / Петренко Л.Р., Андрієнко В.О., Рідей Н.М. // Відтворення родючості ґрунтів у ґрунтозахисному землеробстві / НАУ. – К.: Оранта. – 1998. – 680 с.
14. Пилипенко О.І. Системи захисту ґрунтів від ерозії / Пилипенко О.І., Юхновський В.Ю., Ведмідь М.М. – К.: Златояр, 2004. – 435 с.
15. Полупан М.І. Визначник еколого-генетичного статусу та родючості ґрунтів України / М.Ы Полупан, В.Б. Соловей, В.І. Кисіль, В.А. Величко. К.: Колообіг. – 2005. – 304 с.
16. Полупан М.І. Природний механізм захисту схилових ґрунтів від водної ерозії / Полупан М.І., Балюк С.А., Соловей В.Б., Величко В.А., Волков П.О. – К.: Фенікс, 2011. – 144 с

17.Почвоведение. Учеб. для ун-тов. В 2 ч. // Под ред. В.А. Ковды, Б.Г.Розанова. 4.1. Почва и почвообразование / Г.Д. Белицкая, В.Д. Васильевская, Л.А. Гришина и др. – М.: Высшая школа. – 1988. – 400 с.

18. Розанов Б.Г. Морфология почв / Б.Г. Розанов. – М.: Изд.-во МГУ. – 1983. – 320 с.

19. Туренко А.М. Почвенно-географическое районирование, генезис, свойства и классификация почв Карпатской горной провинции / А.М. Туренко // Позвозащитная бесплужная обработка почв / Сб. науч. тр. - К.: Изд-во УСХА. - 1991. – С. 29-43.

20. Шикула М.К. Охорона ґрунтів / М.К. Шикула, О.Ф. Гнатенко, Л.Р. Петренко, М.В. Капштик. – К.: «Знання» КОО. – 2004. – 398 с.

21.Berezhnyak E. Changes of Fen Soils after Drainage and Agricultural Use in the Trubizh-Region (Ukraine) / E. Berezhnyak, J. Zeitz // ICID 21st European Regional Conference Frankfurt (Oder) and Slubice, 15-19 May 2005), Germany and Poland. – 2005. – pp. 378–382.

22.Berezhniak E. Environmental stability of Chernozem Soils of the Forest Steppe zone of Ukraine / Iev. Berezhniak, I. Shevchenko, M. Berezhniak // 3rd International Scientific and Practical Conference “Science and Society”. Vol. 1. 20-21 March 2013, London, Berforts Information Press Ltd., Stevenage, UK – 2013. – P. 202-213. http://scieuro.com/category/articles

23. Berezhniak E. Minimization of Water Erosion Degradation Processes on Chernozems in The Forest Steppe Zone of Ukraine / Iev. Berezhniak // The 3rd International Geography Symposium KEMER /Antalya/ TURKEY PROCEEDINGS. – P.33-41.
http://web.deu.edu.tr/geomed/proceedings/download/004_GeoMed_2013_Proceedings_33-41.pdf

24. Black C.A. Soil Fertility Evaluation and Control Lewis Publishers / C.A. Black. – London. – Tokyo. – 1992. – 756 p.

25. Bohn H.Z. Soil Chemistry / H.Z. Bohn, B.Z. McNeal, and G.A. O’Connor. – 2nd Ed. John Wiley & Sons. N.Y. – 1985. – 341 p.

26. Focus on Earth Science. Edited by M.S. Bishop, et al. A Bell and Hovell Company. Columbus, Ohio, Toronto, London, Sydney. 1981. – 536 p.

27. Foth H.D. Fundamentals of Soil Science / H.D. Foth. – 8th Ed. John Wiley & Sons. N.Y. – 1990. – 360 p.

28. Gnatenko O.F. Soil Genesis, Classification, Nomenclature, and Diagnostics / O.F. Gnatenko, L.R. Petrenko, Yu.S. Kravchenko and M.I. Baydyuk. – Kyiv. NAU Publishing Center. 2006. – 50 p.

29. Gnatenko O.F. Soil Science with the Elements of Geology / O.F. Gnatenko, L.R. Petrenko L.R. and M.V. Kapshtyk. – A Methodological Manual for the Students of Agronomy. Kiev. NAU Publication. 1997. – 78 p.

30. Kabisch K. Ponds and Pools-Oases in the Landscape / K. Kabisch and Hemmefrling J. – Edition Leipzig. – 1982. – p. 13-18.

31. Matthews W.H. Geology Made Simple. Made Simple Books / W.H. Matthews. – Doubleday and Company, Inc., Garden City, New York. 1967. – 192 p.

32. McBride M.B. Environmental Chemistry of Soils. Oxford University Press / M.B. McBride. – New York-Oxford. – 1994. – 406 p.

33. Minerals in Soil Environments. Second Edition. Co-Editors: J.B. Dixon and S.B. Weed. #1 in the Soil Science Society of America Book Series. Published by: Soil Science Society of America. Madison, Wisconsin, USA. 1989. – 1244 p.

34. Petrenko L. Fundamentals of Soil Science / L. Petrenko, M. Berezhnyak, T. Dudar, Ye. Berezhnyak. – K.: NAU-druk Publishing House, 2010. – 455 p.

35. Petrenko L.R. Methodological Instruction Book on the Elements of Geology for the Students Majoring in Agronomy / L.R. Petrenko, Yu.S. Kravchenko and V.M. Starodubtzev. – Kyiv. NAU Publishing Center. – 2005. – 96 p.

36. Petrenko L. Soil Science. Practical Methods Manual: methodological Instruction Book for bachelor students of higher educational institutions major in 6.090101 / L. Petrenko, M. Berezhnyak, V. Degtyarov, Yu. Kravchenko, O. Tonkha, Ye. Berezhnyak, O. Bykova. – K.: Publishing OOO “NPP “Interservice”, 2014. – 428 p.

37. Petrenko L.R. Soil Science with Elements of Geology. Methodological Guidelines for the Laboratory Classes / L.R. Petrenko, V.M. Starodubtzev and Y.S. Kravchenko. – Kyiv. NAU Publishing Center. 2005. – 62 p.

38. Starodubzev V.M. / V.M. Starodubzev, O.A. Kolodyazhny, L.R. Petrenko et al. Soil Cover and Land Use in Ukraine. Edited by A.G.Tarariko. Kyiv. Nora-Print. 2000. – 96 p.

39. Soil Biology: Effects on Soil Quality. Edited by J.L.Hatfield and B.A. Stewart. Lewis Publishers. – London-Tokyo. – 1994. – 169 p.

40. Troech F.R. Soil and Water Conservation / F.R. Troech, J.A. Hobbs and R.L.Donahue. – 2nd Ed. Prentice Hall, New Jersey. – 1991. – 530 p.

GLOSSARY

The glossary was compiled from three sources of geologic and soil science terms with Ukrainian remarks:

1. M.S. Bishop, B.S. Sutherland, and P.G. Lewis. Focus on Earth Science. C.E. Merril Publishing Co. A Bell and Howell Company. Columbus. Ohio; Toronto, London, Sydney, 1981, pp. 523-530.
2. A.Maton, J.Hopkins, S.Johnson, et. al. Dynamic Earth. Prentice Hall. Englewood Cliffs, New Jersey Needham, Massachusetts. 1993, pp. 171-173.
3. H.D. Foth. Fundamentals of soil science.-8th ed. Published by John Wiley & Sons, Inc. 1990, pp. 342-352.

We are thankful to the authors.

A

A horizons (поверхневі мінеральні горизонти). Mineral horizon that formed at the surface, or below an O horizon (поверхневий органогенний горизонт, як лісова підстилка чи степовий войлок), and are characterized by an accumulation of humified OM intimately mixed with the mineral fraction.

Abrasion (абразія). The wearing away of a substance by solid particles carried by wind, water, or other forces.

Abyss (абісальна зона; abyss = прірва). The deepest part of the ocean floor.

Acid Soil (кислий ґрунт). Soil with a pH value < 7.

Acidic cations (катіони-носії кислотності ґрунту). Hydrogen ions or cations that, on being added to water, undergo hydrolysis, resulting in an acidic solution. Examples in soils are H^+, Al^{3+}, and Fe^{3+}.

Acidity, residual (кислотність ґрунту залишкова). Soil acidity that is neutralized by lime or other alkaline materials, but that cannot be replaced by an unbuffered salt solution.

Acidity, salt-replacable (кислотність ґрунту, що витісняється розчином солі, обмінна кислотність). The aluminum and hydrogen that can be replaced from an acid soil by an unbuffered salt solution such as KCl. Essentially, the sum of the exchangeable Al and H.

Acidity, total (кислотність ґрунту повна). Sum of salt-replacable and residual acidity.

Acquic (тип водного режиму, приблизно відповідає застійному). A mostly reducing soil moisture regime nearly free of dissolved oxygen due to saturation by ground-water or its capillary fringe and occurring at periods when the soil temperature at 50 centimeters below the surface is above 5°C.

Activity (активність іонів у розчині). Informally, may be taken as the effective concentration of a substance in a solution.

Aerate (забезпечувати аерацією). To allow or promote exchange of soil gases with atmospheric gases.

Aeration porosity (пористість аерації). The fraction of the bulk soil volume that is filled with air at any given time, or under a given condition, such as specified soil-water matric potential.

Aggregate (структурний агрегат ґрунту). A unit of soil structure, usually formed by natural processes, and generally below 10 millimeters in diameter.

Air dry (повітряно сухий). The state of dryness at equilibrium with the water content in the surrounding atmosphere.

Albic horizon (горизонт альбік, елювіальний горизонт E ґрунтів різної генези). A mineral soil horizon from which clay and free iron oxides have been removed or segregated. The color of the horizon is determined primarily by the color of primary sand or silt particles rather than coatings on these particles. An E horizon.

Alfisols (альфі солі, найкраще відповідають ясно-сірим та сірим лісовим ґрунтам України, ґрунтовий порядок американської класифікації). Mineral soil that have umbric or orchic epipedons, argillic or kandic horizons, and that have plant-available water during at least 90 days when the soil is warm enough for plants to grow. Alfisols have a mean annual soil temperature below 8°C or a base saturation in the lower part of the argillic horizon of 35 percent or more when measured at pH 8.2.

Alkaline soil (лужний ґрунт). Any soil having a pH above 7.

Alluvial (алювіальний). Deposits of sand, mud, and other materials transported by flowing water.

Alluvial fan (алювіальний фен; віяло-подібні відклади; fan = віяло). A fan-shaped deposit of sediments formed at the point where a river leaves the mountains and runs out onto a plain.

Ammonification (амоніфікація). The biological process leading to the formation of ammoniacal nitrogen from nitrogen-containing organic compounds.

Ammonite (амоніт). Flat spiral fossil shell of a group of invertebrates abundant in the Mesozoic era.

Ammonium fixation (фіксація амонію). The process of converting exchangeable or soluble ammonium ions to those occupying interlayer positions to K^+ in micas.

Amorphous material (аморфний матеріал). Noncrystalline soil constituents.

Amphibole (амфібол, рогова обманка). One of the rock-forming mineral groups containing iron, magnesium, calcium, and aluminum silicates.

Andisols (андисолі, вулканогенні ґрунти). Mineral soils developed in volcanic ejecta that have andic soil properties. A soil order.

Angiosperm (покритонасіннева рослина). Flowering plant having seeds in a protective covering which enables them to survive and reproduce.

Anion Exchange capacity (AEC, ємність обміну аніонів). The sum total of exchangeable anions that a soil can adsorb.

Anticline (антикліналь). Upward fold or bend of rock strata.

Anticyclone (антициклон). System of wind rotating about a center of high atmospheric pressure, turning clockwise in the northern hemisphere and counterclockwise in the southern hemisphere.

Aquifer (водоносний шар породи). Water-bearing rock.

Argillic horizon (ілювіальний горизонт аргілік). A mineral soil horizon that is characterized by the illuvial accumulation of layer-silicate clays.

Arid climate (посушливий клімат). Climate having insufficient rainfall for vegetation.

Aridic (арідний посушливий, тут водний режим ґрунту). A soil moisture regime that has no plant available water for more than half the cumulative time that the soil temperature at 50 centimeters below the surface is above 5°C and has no period as long as 90 consecutive days when there is water for plants, while the soil temperature 50 centimeters is continuously above 8°C.

Aridisols (Арідісолі, ґрунти з посушливим водним режимом). Mineral soils that have an aridic moisture regime but no oxic horizon. A soil order.

Artesian water (артезіанська вода). Ground water under enough pressure to rise above the aquifer containing it.

Asteroid (астероїд). Fragment of material similar to planetary matter which orbits the sun between Mars and Jupiter.

Asthenosphere (астеносфера). Zone in the mantle where the material is near its melting point and acts almost like fluid.

Astronomical unit (астрономічна одиниця відстані). Average distance between the Earth and the Sun, equal to 49,600,000 km.

Atmosphere (атмосфера). Gaseous mass surrounding the Earth or other celestial bodies.

Atoll (атол). Ring-shaped coral reef often surrounding a body of water.

Available nutrients (доступні рослинам елементи живлення). Nutrient ions or compounds in forms that plants can absorb and utilize in growth.

Available water (доступна рослинам вода у ґрунті). The portion of water in a soil that can be absorbed by plant roots; the amount of water between in situ field capacity (FC) and the permanent wilting point (PWP).

B

B horizons (перехідні горизонти). Horizons that formed below an A, E, or O horizon that have properties different from the overlying and underlying horizons owing to the soil forming processes.

Bar (одиниця тиску або всмоктувальної сили ґрунту по відношенню до води). Aunit of pressure equal to million dynes per square centimeter.

Basic cation saturation percentage (ступінь насичення ґрунту основами, але, на відміну від української практики, виміряний при pH 7 чи 8.2). The extent to which the cation exchange capacity (CEC) is saturated with alkali (sodium and potassium) and alkaline earth (calcium and magnesium) cations expressed as a percentage of the cation exchange capacity, as measured at a particular pH, such as 7 or 8.2.

Batholith (батоліт). Large shield-shaped mass of intensive igneous rock extending to unknown depths.

Bedrock (підстиляюча, коріння порода). The layer of rock beneath the soil.

Benthos (бентол). Organisms that live on the bottom of the ocean.

Big Bang Theory (теорія великого поштовху). A theory that the universe originated from the explosion of a huge mass of matter and that the pieces are still flying apart.

Biomass (біомаса). The total mass of living microorganisms in a given volume or mass of oil.

Biosequence (біосукцесія, еволюційна послідовність). A sequence of related soils, differing one from the other, primarily because of differences in kinds and numbers of plants and soil organisms as a soil-forming factor.

Black hole (чорна діра). Collapsed mass of extremely high density matter in outer space. Density is so great that neither light no particles can get out its vicinity.

Brachiopol (брахіопод). Phylum of marine invertebrates having two unequal shells or valves.

Buffer power (буферність). The ability of ions associated with the solid phase buffer changes in ion concentration in the solution phase.

Bulk density soil (щільність, щільність складення, об’ємна маса ґрунту). The mass of dry soil per unit bulk volume, expressed as grams per cubic centimeter, g/cm^3. The bulk volume is determined before drying to constant weight at 105°C.

Buoyance (виштовхувальна сила). Resultant or upward forces exerted by a fluid on a submerged or floating body, equal to the mass of the fluid displaced by the body.

Butte (бут). Tower-like hill or ridge which rises abruptly from and stands above the surrounding plain.

C

Calcareous soil (карбонатний ґрунт). Soil containing sufficient free $CaCO_3$, and/or, $MgCO_3$ to effervesce visibly when treated with cold 0.1 molar HCl. These soils usually contain from as little as 1 to 20 percent $CaCO_3$ equivalent.

Calcic horizon (горизонт кальцик, дуже карбонатний горизонт). A mineral soil horizon of secondary carbonate enrichment that is more than 15 centimeters thick, has a $CaCO_3$ equivalent above 150 g/kg, and has at least 50 g/kg more $CaCO_3$ equivalent that the underlying C horizon, наприклад горизонт скупчення білозірки в Українських степових ґрунтах (chernozem calcic, kastanozem calcic).

Caldera (кальдера, старе широке жерло вулкана). A roughly circular, steep-sided pit at the top of a volcanic cone whose diameter is at least three times its depth.

Cambic horizon (горизонт камбік). A mineral soil horizon that has a texture of loamy very fine sand or finer, has soil structure rather than rock structure, contains some weatherable minerals, and is characterized by alteration or removal of mineral material, or the removal of carbonates (бурі лісові ґрунти гірських регіонів України за номенклатурного ФАО названі камбісолями).

Capillary fringe (капілярна торочка). A zone in the soil just above the water table that remains saturated or almost saturated with water.

Carbonation (карбонатизація). The process in which carbonic acid reacts chemically with other substances.

Carbon-organic nitrogen ratio (відношення *C/N*). The ratio of the mass of organic carbon to the mass of organic nitrogen in soil, organic material, plants, or the cells of microorganisms.

Carnivorous (м'ясоїдний). Flesh-eating.

Cast (відбиток). Mineral or mud deposited and hardened where the body of an organism decayed.

Cat clay (кисла глина). Wet clay soils containing ferrous sulfide, which become highly acidic when drained.

Catena (катена, послідовність ґрунтового покриву на спряжених елементах рельєфу). A sequence of soils of of about the same age, derived from similar conditions, but having different characteristics because of variations in relief and in drainage.

Cation exchange (обмін катіонів). The interchange between a cation in solution and another cation on the surface of any negatively charged material such as clay colloid or organic colloid.

Cation exchange capacity (ємність обміну катіонів). The sum of exchangeable cations that a soil can adsorb at a specific pH; commonly expressed as milliequivalents per 100 grams or centimoles per kilogram.

Celestial sphere (небесна сфера). Glove with the Earth as center on which movements of stars and planets are plotted or mapped.

Chemical rock (хімічний осад). A non-clastic sedimentary rock formed by inorganic processes such as evaporation.

Chemical weathering (хімічне вивітрювання). Weathering that involves changes in the chemical makeup of rocks.

Chroma (інтенсивність забарвлення ґрунту). The relative purity, strength, or saturation of a color, directly related to the dominance of the determining wave length of the light and inversely related to grayness; one of the three variables of color.

Chromosphere (хромосфера). Layer of hot gas surrounding the photosphere of the Sun.

Chronosequence (часова (вікова) послідовність споріднених ґрунтів). A sequence of related soils that differ one from the other, in certain in properties, primarily as a result of time as a soil-forming factor.

Cinder (вулканічний попіл). A small, rough volcanic bomb no more than several centimeters across.

Cinder cone (кратер з вулканічного попелу). A volcano made mostly of cinders and other rock particles that have been blown into the air.

Cirque (цирк). Deep, steep-sided recess or hollow in a mountain caused by glacial erosion.

Clastic, clastic rock (уламкова осадова порода). A sedimentary rock formed from fragments of previously existing rocks.

Clay (глина, мул). (1) A soil separate consisting of particles < 0.002 millimeters in equivalent diameter. (2) A textural class.

Clay films (глинисті (колоїдні) плівки). Coating of clay on the surfaces of soil peds, mineral drains, and in soil pores. (Also called clay skins, clay flows, or agrillans).

Clay mineral (глинний мінерал). Any crystalline inorganic substances of clay size.

Claypan (глинистий ілювіальний горизонт, дуже щільний і непроникний). A dense, compact layer in the subsoil having a much higher clay content than the overlying material, from which it is separated by a sharply defined boundary. Claypans usually impede the movement of water and air and the growth of plant roots.

Cleavage (злам). Tendency in a rock or mineral to split along planes determined by the crystal structure of a mineral; the tendency of a mineral to break along, definite surfaces.

Climate (клімат). Average of local temperature, precipitation, and wind conditions over a period of years.

Colluvium (колювій). A general term applied to deposits on a slope or at the foot of a slope or cliff that were moved there chiefly by gravity.

Comet (комета). Independent mass of cosmic dust and ice which orbits the Sun.

Compaction (ущільнення). Decrease in volume of sediments due to pressure of overlying material or drying.

Composite volcano (складний вулкан). A volcano built of alternating layers of rock particles and lava.

Compound (сполука). Two or more elements chemically joined.

Compression (компресія). The type of stress that squeezes rocks together.

Concave (увігнутий). Hollowed or rounded inward like the inside of a bowl.

Conchoidal (раковистий (злам)). Curved like the inside of a shell.

Concretion (конкреція). A local concentration of a chemical compound, such as calcium carbonate or iron oxide, in the form of a grain or nodule of varying size, shape, hardness, and color.

Conduction (провідність). Movement of heat, light, electricity, or sound by contact among particles.

Conglomerate (конгломерат). Rock consisting of rounded and water-worn pebble sized fragments by another mineral.

Consistency (консистенція). The manifestation of the forces of cohesion and adhesion acting within the soil at various water contents, as expressed by the relative ease with which a soil can be deformed or ruptured.

Consumptive water use (сумарне водоспоживання культур). The water used by plants in transpiration and growth, plus vapor loss from adjacent soil or show, or from intercepted precipitation in any specified time. Usually expressed as equivalent depth of free water per unit of time.

Contact metamorphism (контактний метаморфізм). Change in rock resulting from contact with hot magma or igneous rock.

Continental drift (континентальний дрейф). Slow motion of continents and sea floor adjacent to them.

Continental shield (континентальний щит). Region where the oldest rocks of the continent are exposed at the surface.

Contour interval (інтервал контуру). A distance between contour lines on a topographic map.

Contour lines (контурні лінії). Lines drawn on a topographic map to join all points of the same elevation.

Convection (конвекція). Movement of heat and matter because of differences in density usually resulting from differences in temperature (or pressure).

Convergent boundary (межа між плитами, що зближуються). A plate boundary at which plates come together.

Convex (опуклий). Curved or rounded out ward like the outside of a bowl.

Coriolis force (сила Кореоліса). Apparent force due to the Earth's rotation that causes a body in motion to be deflected from its initial path.

Corona (тут сонячна корона). Low-density cloud of gases surrounding the Sun. Brightness I about one millionth of the Sun's brightness.

Cosmic (космічний). Pertaining to the part of the universe beyond the Earth's atmosphere.

Cosmic ray (космічний промінь). An extremely high speed ion; an atomic nucleus. Solar cosmic rays are blown outward from the Sun; galactic cosmic rays arrive from any direction. All elections are removed from cosmic rays.

Crater (кратер). A funnel-shaped pit at the top of a volcanic cone whose diameter is less than three times its depth.

Creep (зсуви). Slow mass movement of soil and soil material down relatively steep slopes primarily under the influence of gravity.

Crest (кора). Highest natural point of a hill, mountain, anticline, or wave.

Crust (кірка). A soil-surface layer ranging in thickness from a few millimeters to a few tens of millimeters, which, when dry, is much more compact, hard, and brittle than the material immediately beneath it.

Cryic (країк, тип теплового режиму ґрунту). A soil temperature regime that has mean annual soil temperatures of above 0°C but below 8°C at 50 centimeters.

There is more than a 5°C difference between mean summer and mean winter soil temperatures.

Crystal (кристал). (1) A solid in which the atoms or molecules are arranged in a definite pattern that is repeated over and over again. (2) Solid body bounded by plane surfaces showing a regularly repeated arrangement of atoms.

Crystal faces (грані кристалу). Flat surfaces of a crystal which join at well defined angles.

Cyclone (циклон). System of wind rotating about a center of low atmospheric pressure, turning clockwise, in the southern hemisphere and counterclockwise in the northern hemisphere.

D

Darcy's law (закон Дарсі). A law describing the rate of flow of water through porous media. Named for Henry Darcy of Paris, who formulated it in 1856 from extensive work on the flow of water through sand filter beds.

Daughter element (дочірній елемент). Element formed from another by radioactive decay.

Decomposition (розклад). Chemical separation of minerals and rocks into elements or simpler compounds.

Deflation (дефляція). Removal by wind of loose material from the land surface.

Deformation (деформація). In geology, any change in the original shape or volume of rocks.

Delta (дельта). (1) A triangular formation of sediments deposited at the mouth of a large river that flows into a lake or ocean. (2) Triangular shaped alluvial deposits at the mouth of a river.

Dendritic (деревоподібний). Branching figure resembling a tree.

Denitrification (денітріфікація). Reduction of nitrate to nitrite and molecular nitrogen or nitrogen oxides by microbial activity or by chemical reactions involving nitrites.

Density (щільність). The amount of matter in a given space; the mass per unit volume.

Density current (течія, що виникає через різницю у щільності води). Ocean current caused by differences in density due to unequal amounts of suspended or dissolved substances in water or due to differences in temperature.

Deposition (відкладення). (1) The process by which sediments are laid down in new locations. (2) Laying down of possible rock-forming sediments; (3) precipitation of dissolved substance.

Dew point (точка роси). Temperature at which relative humidity is 100 percent and water vapor condenses.

Diatoms (діатомові водорості). One-celled alga which has crystal-like covering of silica.

Diffuse double layer (дифузна частина шару іонів різного знаку, що оточує колоїдну частку). A heterogeneous system, that consist of a solid surface having a net electrical charge, together with an ionic swarm under the influence of the solid and a solution phase that is in direct contact with the surface.

Dike (дайк). Body of igneous rock flat cuts across another rock body.

Dinosaur (динозавр). Group of large extinct reptiles.

Disintegration (розпал на частки). Mechanical breaking of material into small particles during weathering.

Displacement (заміщення, зміщення). Relative movement of two sides of a fault measured in any specified direction.

Divergent boundary (границі між плитами, що розходяться). A plate boundary at which plates move apart.

Doldrums (затишшя). Calm, windless area near the equator.

Dome (купол). A raised area shaped roughly like the top half of a sphere, often formed by magma pushing upward on the rock layers above it.

Doppler effect (ефект Допплера). Change in the frequency with which waves from a given source reach the observer when the source and the observer are in rapid motion with respect to one another, also known as the red shift.

Drainage basin (басейн головної річки). The area drained by a main river and its channels.

Drainage system (гідрографічна мережа). The network of streams and other bodies of running water that ultimately drains into an area’s mass river.

Drumlin (друмлін). An oval-shaped mound of till.

Duripan (дурінен, зцементований горизонт у профілі ґрунту). A mineral soil horizon that is cemented by silica to the point that air-dry fragments will not shake in water or HCl.

E

E horizon (елювіальні горизонти). Mineral horizons in which the main feature is loss of silicate clay, iron, aluminum, or some combination of these, leaving a concentration of sand and silt particles of quartz or other resistant minerals.

Earthquake (землетрус). The shaking and trembling that result from the sudden movement of part of the Earth’s crust.

EC (електропровідність, ЕП). The electrolytic conductivity of an extract from saturated soil, normally expressed in the units of decisiveness per water at 25°C.

Eclipse (затемнення). Passing of a luminous body into the shadow of another body.

Ecliptic (екліптика). Line in the sky along which the Sun appears to move eastward 360° during the year; this line represents the plane of the Earth’s orbit.

Ecology (екологія). A study of the relationships between organisms and their environment.

Ectomycorrhiza (ектомікориза, грибокорінь зовнішній). A mycorrhizal association in which the fungal micellia extend inward, between root cortical cells, to form a network and outward into the surrounding soil.

Edaphology (едафологія). The science that deals with the influence of soil on living things, particularly plants.

Eddy (вихор). Current of water, air, or gas running contrary to the main current.

Electron cloud (електронна хмара). Portion of an atom consisting of negative electricity surrounding the nucleus.

Element (хім. елемент). Any substance which, in its pure form, cannot be separated into simpler substances.

Elevation (підвищення). Height above sea level.

Ellipse (еліпс). Curved plane surface generated by a point that moves so the sum of its distance from two fixed points is constant.

Eluviation (вимивання, елювіація). The removal of soil material in suspension or in solution from a layer or layers of soil.

Endomycorrhiza (ендомікориза, грибокорінь внутрішній). A mycorrhizal association with intracellular penetration of the host root cortical cells by the fungus as well as outward extension into surrounding soil.

Energy (енергія). Capacity for doing work.

Energy level (енергетичний рівень). One of a series of levels in which elections vibrate around the nucleus of an atom.

Entisols (ентісолі, ґрунтовий порядок). Mineral soils which have no distinct subsurface diagnostic horizons within 1 meter of the soil surface. A soil order.

Environment (середовище). Sum total of all conditions surrounding an organism or community.

Epicenter (епіцентр). Point on the Earth's surface directly above the focus of an earthquake.

Equinox (рівнодення). Position where the center of the sun crosses the plane of the Earth's equator or making day and night of equal length.

Erosion (ерозія ґрунту). (1) The wearing away of the land surface by running water, wind, ice, and other geologic agents, including such processes as gravitational creep. (2) Detachment and movement of soil or rock by water, wind, ice, or gravity.

Erosion (ерозія). (1) Process by which materials of the Earth's surface are loosened or dissolved and removed. (2) The process by which the products of weathering are removed from one place to another.

Erosion potential (кількісний показник податливості ґрунту до ерозії, ерозійний потенціал). A numerical value expressing the inherent erodibility of a soil.

Esker (ескер). Serpentine ridge or hill of sand and gravel deposited within stream channels in decaying glacier ice sheet.

Eutrophic (евтрофічний, що стосується евтрифікації). Having concentration of nutrients optimal, or nearly so for plant or animal growth.

Evaporation (випаровування). Physical change from liquid to gas; process by which water becomes a vapor at a temperature below the boiling point.

Evaporite (евапорит). Product of evaporation; sediment left after evaporation of a solvent.

Evapotransporation (евапотранспірація, транспірація рослин плюс випаровування з поверхні ґрунту). The combined loss of water from a given area, and during a specified period of time, by evaporation from the soil surface and by transpiration from plants.

Exchangeable cation percentage (ступінь насичення (%) ГВК певним катіоном, наприклад натрієм (ESP, exchangeable sodium percent), кальцієм, тощо). The extent to which the adsorption complex of a soil is occupied by a particular cation.

Exchangeable ion (обмінний іон). A cation or anion held on or near the surface of a solid particle, which may be replaced by other ions of similar charge that are in solution.

Exfoliation (ексфоліація, відлущення шарів твердої породи). The breaking off of curved sheets or slabs parallel to a rock's surface due to weathering (наше: десквамація).

Exosphere (екзосфера). Outer layer of the Earth's atmosphere which contains helium, hydrogen, radioactive particles, and bands of radiation.

Extinct (вимерлий). No longer existing.

Extrusive (ефузивний). Igneous rock which have been consolidated at a near the Earth's surface; an igneous rock formed from lava.

F

Facies fossil (обмежений, фаціальний викопний вид організмів). A species of fossil adapted to life in a limited environment.

Fallowing (утримання ґрунту під паром). The practice of leaving land uncropped for periods of time to accumulate and retain water and mineralized nutrient element.

Fault (диз'юнктивна дислокація). A break or crack along which rocks move; crack in the Earth's crust where surface rocks have slipped up, down, or sideways.

Fault-back mountain (гора-горст). A mountain formed by blocks of rock uplifted from normal faults.

Fertigation (фертигація). Application of plant nutrients in irrigation water.

Fertility, soil (родючість ґрунту). The ability of a soil to supply elements essential for plant growth without a toxic concentration of any element.

Fertilizer (добриво). Any organic or inorganic material of natural or synthetic origin (other than liming material) that is added to a soil to supply one or more elements essential to the growth of plants.

Fibric material (волокнистий матеріал). Mostly undercomposed plant remains that contain large amounts of well-preserved and recognizable fibers.

Field capacity, in situ (field water capacity) (польова, або найменша вологоємність, НВ). The content of water, on mass or volume basis, remaining in a soil 2 or 3 days after having been wetted with water and after free drainage is negligible.

Film water (плівкова вода). A thin layer of water, in close proximity to soil-particle surfaces, that varies in thickness from 1 or 2 to perhaps 100 or more molecular layers.

Fission (розщеплення). A splitting or breaking up into parts, particularly the splitting of an atomic nucleus resulting in the release of large amounts of energy.

Fixation (фіксація). The process by which available plant nutrients are renderedless available or unavailable in the soil.

Flood plain (заплава). A flat area that is found on both sides of a river or stream and is formed by sediments deposited during floods.

Floodplain (заплава (річки)). The land bordering a stream, built up of sediments from overflow of the stream, and subject to inundation when the stream is at flood stage.

Flux (потік (речовини чи енергії)). The time rate of transport of a quantity across a given area.

Focus (центр, фокус). The underground point of origin of an earthquake, where the rocks break and move.

Fold (складка). (1) A bend in a rock. (2) A bend in rock strata.

Foliation (листоподібна структура, сланцева структура). Structure in certain metamorphic rocks resulting from segregation of different minerals into parallel layers.

Foot wall (стіна зсуву). The block of rock below a fault.

Fossil (викопний об'єкт). (1) The preserved remains or traces of an ancient organism. (2) Record of past life, such as a shell, bone, or impression, preserved in the Earth’s crust.

Fracture (злам). (1) In minerals, the way a mineral that does not cleave breaks along a rough or jagged surface. (2) Distinctive manner of breaking in a mineral other than along a plane surface.

Fracture (розлом). A break or a crack.

Freezing point (температура замерзання). A temperature at which a liquid becomes a solid.

Frequency (частота). Number of repetitions of a periodic wave per unit of time.

Friable (ламкий, крихкий). A consistency term pertaining to the ease crumbing of soils.

Front (атмосферний фронт). Weather term for boundary between different air masses.

Frost action (розтріскування через замерзання). The breaking apart of a rock caused by the water freezing and expanding within cracks.

Fulvic acid (фульвокислота). The colored material, that remains in solution after the removal of humic acid by acidification.

Fusion (злиття). Union by melting; particularly union of atomic nuclei to form heavier nuclei resulting in the release of enormous quantities of energy.

G

Galaxy (галактика). System or community of stars.

Gem. Precious or semiprecious stone which may be polished for an ornament (gemstone).

Geode (жеода). Hollow, globular body of rock often lined with inward growing mineral crystals.

Geologic column (геологічна колонка). An arrangement of rock layers in the order of their deposition.

Geothermal (геотермальний). Energy from deposits of heat in the Earth’s crust.

Geyser [‘gaɪza(r)] (гейзер). Spring that irregularly throws forth jets of hot water and steam.

Gibbous [‘gaɪbes] (про місяць). Moon when more than half but not its entire disc is illuminated (може, щербатий місяць).

Gibbsite (гібсит, мінерал). A mineral with a platy habit that occurs in highly weathered soils. $Al(OH)_3$.

Glacial drift (моренний матеріал). Rock debris that has been transported by glaciers and deposited, either directly from the ice or from the melt water.

Glacier (льодовик). A large mass of moving ice and snow.

Goethite (гетит, мінерал). A yellow-brown iron oxide mineral that is very common and is responsible for the brown color in many soils. $FeOOH$.

Gradient (градієнт). Slope, particularly of a stream or land surface.

Gravitation (гравітація). Mutual attraction between all matters.

Great soil group (велика ґрунтова група). One of the categories in the system of soil classification, that has been used in the United States for many years. Great groups place soils according to soil moisture and temperature, basic cation saturation status, and expression of soil horizons.

Green manure (зелене добриво). Plant material incorporated into soil while green or at maturity, for soil improvement.

Groundwater (ґрунтова вода (у першому від поверхні водоносному горизонті). That portion of the water below the surface of the ground at a pressure equal to or greater than atmospheric.

Gypsic horizon (гіпсоносний горизонт). A mineral soil horizon of secondary calcium sulfate enrichment that is more than 15 centimeters thick.

Gypsum (гіпс). The common name for calcium sulfate $(CaSO_4 \cdot 2H_2O)$, used to supply calcium and sulfur to ameliorate sodic soils.

H

Hachure [ha’tʃuə]. A short dash line at straight angles to contour lines, used for shading to show direction of slope.

Half-life (період напіврозпаду). Time in which half the initial number of atoms of a radioactive element disintegrate into atoms of a daughter element.

Hardness (твердість). Resistance to scratching or abrasion.

Hardpan (ущільнений, зцементований горизонт). A hardened soil layer, in the lower A or in the B horizon, caused by cementation of soil particles with organic matter or with materials such as silica, sesquioxides, or calcium carbonate.

Heavy metals (важкі метали). Those metals that have high density, in agronomic usage includes Cu, Fe, Mn, Mo, Co, Zn, Cd, Hg, Ni, and Pb.

Hematite (гематит, мінерал). A red iron oxide mineral that contributes red color to many soils Fe_2O_3.

Hemic material (органогенний горизонт хемік). An intermediate degree of decomposition, such as two-thirds of the organic material cannot be recognized.

Heterotroph (гетеротроф). An organism capable of deriving carbon and energy for growth and cell synthesis by the utilization of organic compounds.

Hexagonal (гексагональний). Crystal system having three lateral axes intersecting at angles of 60° in one plane and a fourth unequal axis perpendicular to the others.

Histosols (гіпсосолі, торф'яні ґрунти). Organic soils that have organic soil material in more than half of the upper 80 centimeters, or that are of any thickness overlying rock or fragmental materials, which have interstices filled with organic soil materials. A soil order.

Horizons (генетичні горизонти, горизонти материнської породи). (1) Soil layers. (2) Horizons or layers, excluding hard rock, that are little affected by the soil-forming processes.

Hot fluid deposit (гідротермальні відклади). Mineral deposited in joints and fractures of tock by heated water in the crust.

Hue (відтінок; тут – одна з трьох змінних, що характеризують забарвлення ґрунту). One of the three variables of color.

Humic acid (гумінова кислота). The dark-colored organic material that can be extracted from soil by various agents and that is precipitated by acidification to pH 1 or 2.

Humic acid (гумінова кислота). Product of decomposition of plant or animal matter.

Humification (гуміфікація). The process whereby the carbon of organic residues is transformed and converted to humic substances through biochemical or chemical processes.

Humus (гумус). All of the organic compounds in soil exclusive of undecayed plant and animal tissues, their partial decomposition products, and the soil biomass. Resistant to further alteration.

Hurricane (ураган). Storm which develops when warm, moist air carried by trade winds rotates around a low-pressure “eye” as it moves forward.

Hydraulic conductivity (гідравлічна провідність). The proportionality factor in Darcy’s Law as applied to the viscous flow of water in soil.

Hydrocarbon (вуглеводень). Organic compound of carbon and hydrogen.

Hydroelectric power. Energy produced by water power.

Hydrosphere (гідросфера). Water portion of the Earth including water vapor in the air, seas, rivers and groundwater.

Hydroxy-aluminum ($Al(OH)^{2+}$, $Al(OH)_2^+$ та їх сполуки). Aluminum hydroxide compounds of varying composition.

Hyperthermic (тип температурного режиму ґрунту). A soil temperature regime that has mean annual soil temperatures of 22°C or more and a higher than 5°C difference between mean summer and mean winter soil temperatures at 50 centimeters below the surface.

Hypothesis (гіпотеза). Proposition or assumption based on available information offered as an explanation for a problem.

I

Iceberg (айсберг). A part of a glacier that has broken off and drifted into the sea.

Igneous (магматичний). Formed from molten rock.

Illuvial horizon (ілювіальний горизонт). A soil layer or horizon in which material carried from the overlying layer has been precipitated from solution or deposited from suspension. The layer of accumulation.

Illuviation (вмивання і відкладання матеріалу, вимитого зверху). The process of deposition of soil material removed from one horizon to another in the soil, usually from an upper to a lower horizon in the soil profile.

Immature river (незріла (молода) річка). A river in an early stage of development.

Immobilization (іммобілізація). The conversion of an element from the inorganic to the organic form in microbial or plant tissues.

Impermeable (непроникний): having a texture that does not permit liquid to move through the pores

Inceptisols (інсептисолі, ґрунтовий порядок у класифікації ґрунтів США). Mineral soil having one or pedogenic horizons in which mineral materials, other than carbonates or amorphous silica, have been altered or removed but not accumulated to a significant degree.

Index fossil (ключові, визначні, діагностичні, керівні, викопні рештки): fossil with a narrow time range and wide distribution used to identify and date the rock layer in which it occurs

Indicator plants (рослини-індикатори). Plants characteristic of specific soil or site conditions.

Infiltration (інфільтрація). The downward entry of water through the soil surface.

Inorganic (неорганічний). Not formed from living things or the remains of living things.

Intrusive (інтрузивний). Rock which, while fluid, has penetrated into to between other rocks, but has solidified before reaching the surface of the Earth.

Intrusive rock (інтрузивна порода). An igneous rock formed from magma.

Invertebrate (безхребетний). Animal lacking a spinal column.

Ion activity (активність іонів). Informally, the effective concentration of an ion in solution.

Ionosphere (іоносфера). A layer of the Earth’s atmosphere about 80km to 600km above the Earth’s surface containing free electrically charged particles by means of which radio waves are transmitted to distant areas.

Iron oxides (оксиди заліза). Group name for the oxides and hydroxides of iron. Includes the minerals golthite, hematite, ferrihydrite, maghemite, and magnetite. Sometimes referred to as sesquioxides or hydrous oxides.

Iron pan (озалізнений горизонт). An underrated soil horizon in which iron oxide is the principal cementing agent, as in plinthite or laterite.

Ironstone (рудяковий горизонт). Hardened plinthite materials often occurring as nodules and concretions.

Isobar (ізобара). A line on a weather map joining points of equal air pressure.

Isomorphous substitution (ізоморфне заміщення). The replacement of one atom by another of similar size in a crystal structure without disrupting or seriously changing the structure.

Isostasy (рівновага вертикальних сил). The balancing of the downward force of the crust and the upward force of the mantle.

Isotherm (ізотерма). A line on a weather map joining points of equal temperature.

Isotopes (ізотопи). Atoms of the same element having different mass because of the differences in the number of neutrons.

J

Jarosite (ярозит, мінерал). A yellow potassium iron sulfate mineral.

K

Kandic horizon (горизонт кандік). Subsoil diagnostic horizon having a clay increase relative to overlying horizons and having low-activity clays, below 16 meq/100 g or below 16 cmol (+)/kg of clay.

Kaolinite (каолініт). A clay mineral of the kaolin subgroup. It has a 1:1 layer structure and is a nonexpanding clay mineral.

Kettle lake (озеро у котловині (чайнику)). Around, deep lake formed by a huge block of ice left behind by a glacier.

L

Labile (мобільний. легкозмінюваний). A substance that is readily transformed by microorganisms or is readily available to plants.

Laccolith (лаколіт). Mushroom shaped body of intrusive igneous rock which has domed up the overlying rock and has a floor that is usually horizontal in contrast to the larger batholith.

Lagoon (лагуна). Area of shallow salt water possessing a restricted connection with the sea.

Landscape (ландшафт). All the natural features such as fields, hills, forests, water, and such, which distinguish one part of the earth's surface from another part. Usually that portion of the land that the eye can comprehend in a single view.

Landslide (зсув). A large downhill movement of loose rocks and soil caused by the pull of gravity.

Lateral fault (бічний розлом). A fault along which the blocks move horizontally passing each other.

Latitude (широта). Distance north or south of the Earth's equator measured in degrees.

Lattice (решітка, гратка-кристалічна). A regular geometric arrangement of points in aplane or in space. Lattice is used to represent the distribution of replanting atoms or groups of atoms in a crystalline substance.

Lava (лава). (1) Molted rock at the Earth's surface. (2) Fluid rock issuing from a volcano or fissure (тріщина), or same material cooled and solidified.

Leaching (вилуговування, вимивання). The removal of materials in solution from the soil.

Leaching rate (L.R., норма промивання засоленого ґрунту). Термін у глосарії Фота відсутній, але є у підручнику цього ж автора.

Leeward (зустрічний до вітру, завітрений (бік)). Direction toward which the wind is blowing.

Lepidocrocite (лепідокроцит, мінерал). An change iron oxide mineral that is found in mottles and concretions of wet soils. $FeOOH$.

Levee (прирусловий вал). (1) Constructed or natural bank containing a stream channel. (2) In nature, a ridge-like deposit along the sides of a river.

Light-year (світловий рік). Distance that light travels in one year; equal to $9.46*10^{12}$km/yr or 300,000 km/s.

Lime, agricultural (вапняковий матеріал для хімічної меліорації кислих ґрунтів, інколи вживається абревіатура aglime). A soil amendment containing calcium carbonate, magnesium carbonate and other materials, used to neutralize soil acidity and furnish calcium and magnesium for plant growth.

Liquid (рідина). Form of matter without definite shape but having definite volume.

Lithosequence (послідовність одиниць ґрунтового покриву відповідно до нерозривної послідовності їх материнських порід). A group of related soils that differ, one from the other, in certain properties primarily as a result of differences in the parent rock as a soil forming factor.

Lithosphere (літосфера). (1) Solid outermost part of the Earth; the Earth's crust. (2) The topmost solid part of the Earth, which is composed of the crust and some of the mantle. (!) Два визначення, що не зовсім співпадають. Українській терміносистемі відповідає the Earth's crust = земна кора.

Load (твердий стік): the amount of sediment carried by a stream

Loam (суглинок). A soil textural class.

Loess [lɛs, loh-ehs] (лес, пухка пилувата порода). Material transported and deposited by wind and consisting of predominantly silt-sized particles.

Longitude (довгота). Distance east or west of the prime meridian at Greenwich, England, measured in degrees.

Luster (блиск). (1) Character of light reflected by a mineral. (2) The way a mineral reflects light from its surface.

Luxury uptake (надмірне поглинання). The absorption by plants of nutrients in excess of their need for growth.

L-waves (бічні сейсмічні хвилі). Destructive earthquake waves which travel parallel to the Earth's surface.

M

Macronutrient (мікроелемент живлення). A plant nutrient usually attaining a concentration of more than 500 mg/kg (ppm0 in nature plants.

Maghemite (маггеміт, мінерал). A dark, reddish-brown, magnetic iron oxide mineral chemically similar to hematite, but structurally similar to magnetite. Fe_2O_3. Often found in well-drained, highly weathered soils of the tropical regions.

Magma (магма). Molten rock material which is formed beneath the Earth's crust and from which igneous rocks are solidified.

Magnetite (магнетит, мінерал). A black, magnetic iron oxide mineral usually inherited from igneous rocks. Often found in soils as black magnetic sand grains.

Malleable (здатний піддаватись обробці коваля). Capable of being shaped by pounding.

Mammal (ссавець). Any of a class of higher vertebrates that nourish their young with milk secreted from glands.

Manganese oxides (оксиди марганцю). A group term for oxides of manganese. They are typically black and frequently occur as nodules and coatings on ped faces, usually in association with iron oxides.

Mantle (мантія). Layer of the Earth between the crust and the core.

Manure (органічне добриво). The excreta of animals with or without an admixture of bedding or litter, fresh or at various stages of further decomposition or composting.

Mass (маса). Measure of the quantity of matter in a body.

Mass flow (механізм кореневого живлення, масовий потік). The movement of solutes associated with the net movement of water.

Mass number (масове число). Total number of protons and neutrons present in the nucleus of each atom of a given element.

Mass wasting (переміщення матеріалу по схилу). The downhill movement of sediments due to gravity.

Massive (масивний). Mineral form of crystals too small to be seen without a microscope.

Matter (матерія). Anything that has mass and occupies space.

Mature river (зріла ріка). A river that has been developing for many thousands of years.

Meander (меандр). (1) A loop in a river. (2) Turn or loop-like bend in a stream channel.

Mechanical weathering (механічне вивітрювання). Weathering that does not involve changes in the chemical makeup of rock.

Melanic horizon (горизонт меланік). A thick, dark colored surface horizon having andic soil properties.

Melting point (температура плавлення). Temperature at which a solid becomes a liquid.

Meltwater (тала вода). Formed by the water from melting ice or snow; water from melting ice or snow.

Mesa. Isolated hill having steeply sloping sides and a level top protected by a resistant layer of rock.

Mesic (характер теплового режиму ґрунту). A soil temperature regime that has mean annual soil temperatures of 8°C or more but less than 15°C, and more than 5°Cdifference between mean summer and mean winter temperatures at 50 centimeters below the surface (in Ukraine heat regime is evaluated by the sum of temperatures over 10°C at 20 centimeters below the surface).

Mesosphere (мезосфера). Layer of the Earth's atmosphere about 50km to about 85km above the Earth's surface containing ozone which absorbs ultraviolet rays.

Metal (метал). Opaque, fusible, ductile, lustrous element.

Metamorphic (метаморфічний). (1) Changed in form as a result of chemical reactions, heat and/or pressure. (2) Rock changed in composition or texture after consolidation as a result of deformation and/or increased temperature.

Metamorphism (метаморфізм). The process in which metamorphic rock is formed.

Meteor (метеор). Meteoroid that reaches the Earth's atmosphere and burns up.

Meteorite (метеорит). Meteor which reaches the surface of the Earth.

Meteoroid (метеороїд). Fragment of cosmic material too small to be observed from the Earth.

Meteorology (метеорологія). Science dealing with the atmosphere and its phenomena, particularly relating to the weather.

Microclimate (мікроклімат). The sequence of atmospheric changes within a very small region.

Microfauna (мікрофауна). Protozoa, nematodes, and arthropods of microscopic size.

Microflora (мікрофлора). Bacteria (including actinomycetes), fungi, algae, and viruses.

Micronutrient (мікроелемент живлення). A chemical element necessary for plant growth found in small amounts, usually less than 100 mg/kg in the plant. These elements consist of B, Cl, Cu, Fe, Mn, Mo, and Zn.

Mid-ocean ridge (серединно-океанічний хребет). (1) Mountain ridge which extends for about 64,000km roughly parallel to continental margins. (2) An undersea mountain chain where new ocean floor is produced; a constructive (divergent) plate boundary (два визначення доповнюють одне одного).

Migrating dunes (рухомі піски, дюни, що пересуваються). Dunes which move more or less as a unit because of wind action.

Mineral (мінерал). (1) A naturally occurring, inorganic solid that has a definite chemical composition and crystal shape. (2) Inorganic substance which occurs in nature, in a solid state, with a definite chemical composition and characteristic internal atomic pattern.

Mineral soil (мінеральний ґрунт). A soil consisting predominantly of, and having properties determined predominantly by, mineral matter. Usually contains less than 200 g/kg of organic carbon.

Mineralization (мінералізація, повний розклад). The conversion of an element from an organic form to an inorganic state as a result of microbial activity.

Mixture (суміш). Two or more substances combined in any proportion which, unlike a chemical compound, retain their identity and can be separated by mechanical means.

Mohorovicic discontinuity (лінія Мохоровичича). Position within the Earth at which seismograph study indicates an abrupt change in density; a boundary between the solid crust and the plastic mantle often referred to as Moho.

Mold (порожнина після зогнилих решток). A cavity left in a rock after the body of an organism has decayed; a fossil.

Mollisols (молі солі, чорноземи). mineral soils that have a molic epipedon overlying mineral material with a basic cation saturation of 50 percent or more when measured at pH 7.0. A soil order.

Molten (розплавлений). Liquefied by heat.

Montmorillonite (монтморилоніт, мінерал). An aluminum silicate (smectite) with a layer structure composed of two silica tetrahedral sheets and a shared aluminum and magnesium octahedral sheet.

Moon phases (фази місяця). An apparent change in the shape of the Moon's disc because of the Moon's revolution around the Earth and change in its reflected light from the Sun.

Mor (грубий гумус). A type of forest humus in which the A horizon is present and in which there is almost no mixing of surface organic matter with mineral soil.

Moraine (морена). (1) Deposit of unlayered gravel, sand, clay, and boulders left by direct melting of a glacier. (2) A ridge of till left behind by a retreating glacier.

Mottled zone (зона плямистості у профілі ґрунту). A layer that is marked with spots or blotches of different color or shades of color (mottles).

Muck soil (органогенний ґрунт з добре розкладеною органікою). An organic soil in which the plant residues have been altered beyond recognition.

Mulch (мульча). Any material such as straw, sawdust, leaves, plastic film, and loose soil, that is spread upon the soil surface to protect soil and plant roots from the effects of raindrops, soil crusting, freezing, evaporation, and such.

Mulch forming (утримання ґрунту з рештками на поверхні). A system of tillage and planting operations resulting in minimum incorporation of plant residues or other mulch into the soil surface.

Mull (муль, м’який гумус). A type of forest humus in which the O_e horizon may or may not be present and in which there is no O_a horizon. The A horizon consists of an intimate mixture of organic matter and mineral soil with gradual transition between the A horizon and the horizon underneath.

Munsell color system (система Мансела для визначення забарвлення ґрунту). A color designation system that specific the relative degrees of the three simple variables of color: hue, value, and chroma. For example: 10YR 6/4 is a color with a hue = 10YR (yellow red), value = 6, and chroma = 4.

Mutation (мутація). Sudden fundamental change in heredity producing new individuals unlike parents.

Mycorrhiza (мікориза, грибокорінь). Literally “fungus root”. The association, usually symbiotic, of specific fungi with the roots of higher plants.

N

Natural resources (природні ресурси). Materials from the Earth that we can use.

Nebula (туманність). Large cloud of gas and dust in interstellar space which can become a star.

Nekton (нектон). Free-swimming marine organisms.

Nitrogenase (нітрогеназа, фермент). The specific enzyme required for biological dinitrogen fixation.

Normal fault (вид розривної дислокації). A fault in which the hanging wall moves down relative to the foot wall.

No-tillage system (система нульового обробітку ґрунту). A procedure whereby a crop is planted directly into the soil with no preparatory tillage since harvest of the previous crop; usually a special planter is necessary to prepare a narrow, shallow seedbed immediately surrounding the seed being planted.

Nova (така зірка). Star that irregularly, yet suddenly, increases its light output tremendously and then fades into its former obscurity.

Nucleus (ядро). Central point or portion of an atom around which electrons are gathered; central portion of a galaxy.

Nutric horizon (горизонт натрик). A mineral soil horizon that satisfies the requirements of an argillic horizon, but that also has prismatic, columnar, or blocky structure with exchangeable sodium.

Nutrient antagonism (антагонізм елементів живлення). The depressing effect caused by one or more plant nutrients on the uptake and availability of another.

Nutrient interaction (взаємодія елементів живлення). A statistical term used when two or more nutrients are applied together to denote a departure from additive responses occurring when they are applied separately.

Nutrification (нітрифікація). Biological oxidation of ammonium to nitrite and nitrate, or a biologically induced increase in the oxidation state of nitrogen.

O

O horizon (органогенний горизонт підстилки чи войлоку). Layers dominated by organic material, except limnic layers that are organic.

Oasis (оазис). Fertile, green spot in a sandy desert.

Oblate spheroid (просторова форма Землі). Spherical body that bulges at its equator and is flattened at the poles.

Ocean-floor spreading (розширення океанічного дна, розходження плит). The process in which old ocean floor is pushed away from a mid-ocean ridge by the formation of new ocean floor.

Ochric epipedon (горизонт охрик). A thin, light-colored surface horizon of mineral soil.

Octahedron (октаедр). A solid bounded by eight plane faces.

Ooze (відклади океанічного дна). A mud of more than 30 percent organic origin which covers parts of the ocean floor.

Orbit (орбіта). Path of a body in its revolution around another body.

Ore (руда). Mineral or group of minerals which can be mined as a profit.

Organic rock (тут: органогенний осад). A sedimentary rock that is formed either directly or indirectly from material that was once alive.

Organic soil (органогенний ґрунт). A soil that contains a high percentage of organic carbon (> 200 g/kg or > 120-180 g/kg if saturated with water) throughout the solum (частина профілю до материнської породи).

Orthoclase (ортоклаз). Potassium-containing mineral member of the feldspar family; most abundant mineral in granites.

Orthorhombic (ромбічне; про сингонію). Crystal system having three unequal axes intersecting at right angles.

Outwash (відклади води танучих льодовиків). Stratified glacia drift deposited by melt-water streams beyond active glacier ice.

Outwash plain (водно-льодовикова рівнина). Flat, fan-shaped areas in front of terminal moraines formed by sediments deposited by rivers of glacial meltwater.

Oven dry soil (абсолютно сухий ґрунт). Soil that has been dried at 105°C until it reaches constant mass.

Oxbow lake (старичне озеро). A U-shaped lake formed when erosion and deposition cuts off a meander of a river.

Oxidation (окислення). The process in which oxygen chemically combines with another substance (see below).

Oxidation number (ступінь окислення). A number of electrons an atom gains. Looses or shares when it forms chemical bonds. Oxidation is an increase of oxidation number. Oxidation is losing of electrons.

Oxide (оксид). Chemical compound containing oxygen combined with a positive ion or ions.

Oxisols (оксисолі, червоно цвітні ґрунти). Mineral soils that have an toxic horizon within 2 meters of the surface or plant hide as a continuous phase within 30 centimeters of the surface, and that do not have a spodic (ілювіальний) horizon above the oxic horizon. A soil order.

Ozone (озон). Unstable molecule (O_3) formed in the Earth's atmosphere form atomic and molecular oxygen at altitudes from 19km to 35km; molecular oxygen present in the Earth's atmosphere.

P

Paleosol, buried (палеоґрунт, похований ґрунт). A soil formed on a landscape during the geological past and subsequently buried by sedimentation.

Pangaea [pæn'dʒɪə] (Пангея – материк). The single giant land mass that existed more than 200 million years ago and that gave rise to the present-day continents.

Pans (ущільнені, затверділі горизонти, нерідко з високим вмістом мулу). Horizons or layers in soils that are strongly compacted, indurated, or having very high clay content.

Parent material (материнська (ґрунтотворна) порода). The unconsolidated and more or less chemically weathered mineral or organic matter from which the solum of soils is developed by pedogenic processes.

Parsec (парсек). Unit of measure for interstellar space, equal to 3.26 light-years.

Particle density (щільність твердої фази ґрунту). The density of soil particles, the dry mass of the particles being divided by the solid volume of the particles.

Pascal (паскаль). A unit of pressure equal to 1 Newton per square meter.

Peat (торф). Unconsolidated soil material consisting largely of undercomposed, or only slightly decomposed, organic matter accumulated under conditions of excessive moisture.

Ped (пед, грудочка (агрегат) природної структури ґрунту). A unit of soil structure such as an aggregate, crumb, prism, block, or granule, formed by natural process.

Pedon (педон, одиниця ґрунтового тіла). A three-dimensional body of soil with lateral dimensions large enough to permit the study of horizon shapes and relations. Its area ranges from 1 to 10 square meters (m^2).

Penetrability (penetrometer resistance, твердість ґрунту). The ease with which a probe can be pushed into the soil.

Percolation (рух води крізь товщу ґрунту під дією сили гравітації). The downward movement of water through the soil.

Pergelic (тип температурного режиму ґрунту). A soil temperature regime that has mean annual soil temperatures of less than 0°C. Permafrost is present (відповідає тривало-сезонно промиваючому).

Permafrost (горизонт тривалої мерзлоти). A perennially frozen soil horizon.

Permafrost table (глибина сезонного відтанення мерзлотного горизонту). The upper boundary of the permafrost, coincident with the lower limit of seasonal thaw.

Permanent wilting point (PWP, вологість стійкого в'янення ґрунту). The largest water content of a soil at which indicator plants, growing in that soil, wilt and fail to recover when placed in a humid chamber. Often estimated by the water content at-15 bars, -1,500 kilopascals, or -1.5 megapascals soil matric potential.

Petrify (кам'яніти). To change into stone or a stony substance by replacing wood cellulose with silica.

Petrocalcic horizon (щільний, зцементований карбонатами горизонт). A continuous, indurated calcic horizon that is cemented by calcium carbonate and, in some places, with magnesium carbonate.

Petrogypsic horizon (щільний горизонт, зцементований гіпсом). A continuous, strongly cemented, massive gypsic horizon that is cemented with calcium sulfate.

Petroleum (сира нафта). Liquid, flammable hydrocarbon.

pH, soil (pH ґрунту). The negative logarithm of the hydrogen ion activity of a soil.

Phase (тут: фаза ґрунтового покриву, не плутати з фазами ґрунту, як фізичного тіла). A utilitarian grouping of soils defined by soil or environmental features that are not class differentia used in the U.S. system of soil taxonomy, for example, surface texture, surficial rock fragments, salinity, erosion, thickness, and such.

pH-dependent charge (pH-залежний заряд колоїдів ґрунту). The portion of the cation or anion exchange capacity which varies with pH.

Photosphere (фотосфера). Luminous, visible part of the Sun.

Piedmont (п'єдмонд). Are lying along or near the foot of a mountain range.

Placic horizon (горизонт плацик). A black to dark-reddish mineral soil horizon that is usually thin, is commonly cemented with iron, and is slowly permeable or impenetrable to water and roots.

Plagioclase (плагіоклаз). Mineral group containing sodium silicate and/or calcium silicate.

Plankton (планктон). Marine organisms that float in the water.

Plastic soil (пластичний ґрунт). A soil capable of being molded or deformed continuously and permanently, by relatively moderate pressure, into various shapes.

Plate tectonics, plate tectonics theory (тектоніка плит). Explanation of the formation of the Earth’s crustal features which states that the crust is made up of moving plates.

Plateau (плато). Level land rising above adjacent land.

Plow pan (плужна підошва). An induced subsurface soil horizon or layer having a higher bulk density and lower total porosity than the soil material directly above the below, but similar in particle size analysis and chemical properties. The pan is usually found just below the maximum depth of primary tillage and frequently restricts root development and water movement. Also called a pressure pan or plow sole.

Pollution (забруднення). Contamination of the environment with waste.

Polypedon (поліпедон, елементарний ґрунтовий ареал). A group of contiguous similar pedons.

Porphyry (порфір). A rock with distinct crystals in a fine-grained ground mass.

Potassium fixation (фіксація калію у ґрунті). The process of converting exchangeable or water-soluble potassium to that occupying the position of K^+ in the micas.

Precipitate (випадати в осад, осаджувати; тут: хімічний осад). Non-clastic rock which forms when it separates out of chemical solution.

Precipitation (опади). Release of moisture from the atmosphere in the form of rain, snow, sheet, or hail; process of separating minerals from a solution or melt by chemical reaction or evaporation.

Primary mineral (первинний мінерал). A mineral that has not been altered chemically since deposition and crystallization from molten lava.

Prime meridian. Meridian which passes through Greenwich, England, and from which distances east and west on the Earth's surface are measured.

Profile, soil (профіль ґрунту). A vertical section of soil through all its horizons and extending into C horizon.

Prominences (протуберанці сонця). Huge, cloud-like structure of the Sun that appears as great arches.

Pulsar [ʻpu:lsə] (пульсар). A rapidly rotating neutron star that pulsates regularly due to giving off radio waves.

P-wave (вид сейсмічної хвилі). Primary earthquake wave which vibrates back and forth in direction of wave travel.

Pyroxene (піроксен). One of the rock-forming mineral groups containing iron, magnesium, calcium, sodium, and aluminum silicates in which the silicate tetrahedral are arranged in a single chain.

Q

Quasar [ʻkweɪza(r)] (квазар). From quasi-stellar; source of radio energy from space; celestial objects having certain traits of stars, but which are not stars.

R

R layer (горизонт корінної породи). Hard bedrock including granite, basalt, quartzite, and indurated limestone or sandstone that is sufficiently coherent to make hand digging impractical.

Radar [ʻreɪda(r)] (радар). From radio detecting and ranging; instrument for detecting a target and measuring the travel time of a radio pulse sent out from a transmitter and reflected by the target in order to determine the direction of the target.

Radial (радіальний). Stream patterns, streams flowing from a central location.

Radiant energy (променева енергія). Energy which travels in waves.

Radiation (радіація). Process of giving off radiant energy in the form of waves or particles.

Radioactive decay (радіоактивний розпад). Changing one element into another element through loss of charged particles from the atomic nucleus without the influence of heat, pressure, or chemical reaction.

Rainfall erosion index (ерозійний індекс дощу). A measure of the erosive potential of a specific rainfall event.

Rays (промені). Here: bright streaks radiating from some Moon craters.

Reaction (реакція). A chemical change that occurs when two or more substances are mixed, usually in solution.

Reaction, soil (реакція ґрунту). The degree of acidity and alkalinity of a soil, usually expressed as a pH value.

Recrystallization (перекристалізація). Formation of new mineral grains or enlargement of pre-existing mineral grains caused by hot fluid metamorphism.

Recycle (переробляти відходи). To process used materials so they can be reused.

Red shift (червоне зміщення). Shift of all spectral lines in light from receding distant galaxies toward longer wavelengths, or the red end of the spectrum, characteristic of all galaxies; also known as the Doppler effect.

Regolith (реголіт, кора вивітрювання). The unconsolidated mantle of weathered rock and soil material above solid rock.

Relative dates (відносне датування). Dates in the Earth’s history determined with reference to other events.

Relative humidity (відносна вологість). A measure of how much moisture air contains compared to how much it could hold before becoming saturated.

Relief (рельєф). Difference between high and low elevations of a land surface.

Renewable resource (відновний ресурс). Resource which can be replaced.

Reptile (плазун). Cold-blooded vertebrate that moves on its underside or on short legs.

Residual fertility (залишкова родючість). The available nutrient content of a soil carried to subsequent crops.

Reticalate mottling (ретикулярна плямистість). A network of streaks of different color, most commonly found in the deeper profiles of soil containing plinthite.

Retrograde (ретроград). Having direction contrary to general direction of similar bodies.

Revolution (революція, обертання). Moving of a body in a circular course about a central point; a time of major crustal deformations.

Rhizobia (ризобія). Bacteria capable of living symbiotically in roots of leguminous plants, from which they receive energy and often utilize molecular nitrogen.

Rhizosphere (ризосфера). The zone immediately adjacent to plant roots in which the kinds, numbers, or activities of microorganisms differ from that of the bulk soil.

Richter scale (шкала Ріхтера). Scale used to measure the amount of energy released by an earthquake.

Rift zone (рифтова/зона розколу). Central crack in the mid-ocean ridge.

Rill (струмочковий розмив). A small, intermittent watercourse with steep sides; usually only several centimeters deep and, thus, no obstacle to tillage operations.

Rip current (про океанічну течію). Seaward moving current that returns water from the short to greater depths.

Rotation (тут: обертання). Turning motion of a body on its axis.

Runoff (поверхневий стік). That portion of the precipitation on an area which is discharged from the area through stream channels.

S

Saline seep (підтік солей з просо чуваного водою до горизонтів ґрунту). Intermittent or continuous saline water discharge at or near the soil surface under dry-land conditions, which reduces or eliminates crop growth.

Saline soil (засолений ґрунт). A nonsodic soil containing sufficient soluble salt to adversely affect the growth of most crop plants.

Saline-sodic soil (солонцювато-солончаковий ґрунт). A soil containing both sufficient soluble salt and exchangeable sodium to adversely affect crop production under most soil and crop conditions.

Salt balance (баланс солей у ґрунті). The quantity of soluble salt removed from an irrigated area in the drainage water minus that delivered in the irrigation water.

Sand (пісок). 1) A soil particle between 0.05 and 2.0 millimeters in diameter. 2) A soil textural class.

Sand bar (піщана коса, відмілина). A long, underwater ridge of sand.

Sand dune (піщана дюна). A mound of sand deposited by the wind.

Sapric material (саприк, добре розкладений торф). One of the components of organic soils with highly decomposed plant remains. Material is not recognizable and bulk density is low.

Satellite (супутник). Body revolving about a larger body.

Saturation (насичення). Degree to which rock openings are filled with fluid or to which a solution contains all the dissolved material possible at a given temperature.

Saturation extract (витяжка з ґрунту при його повному насиченні водою). The solution extracted from a soil at its saturation water content.

Scale (шкала). On a map, relationship or ratio between distance on the map and distance on the ground.

Scientific theory (наукова теорія). General principle offered to explain observed facts and events.

Sea cave (морська печера). A hollowed out portion of a sea cliff.

Sea cliff (стрімка скеля морського берегу). A steep face of a rock produced by wave action.

Sea stack (морський стовп, колона). A column of resistant rock left behind after a sea cliff has been eroded away.

Secondary mineral (вторинний мінерал). A mineral resulting from the decomposition of a primary mineral or from the reprecipitation of the products of decomposition of a primary mineral.

Sediment (осад). Particles of rock or organic materials that have been carried along and deposited by water and/or wind.

Sedimentary (осадовий). Formed by the compacting and cementing of sediments or by other non-igneous processes at the Earth’s surface.

Sedimentary rock (осадова порода). A rock formed of sediments-either fragments or other rock deposited by wind or water or material precipitated from solutions.

Seismic [ˈsaɪzmɪk] wave (сейсмічна хвиля). An earthquake wave.

Seismograph (сейсмограф). An instrument that detects and measures seismic waves (and records vibrations of the Earth).

Seismologist (сейсмологія). A scientist who studies earthquakes.

Sesquioxides (півтораоксиди). A general term for oxides and hydroxides of iron and aluminum (R_2O_3).

Shadow zone (зона тіні). An area between 103° and 143° on either side of an earthquake focus where no P or S-waves are recorded.

Shearing (деформація). (1) The type of stress that pushes rocks of the crust in two opposite, horizontal directions. (2) Stress resulting from applied forces that cause two adjacent parts of solid to slide <u>past one another</u> parallel to the plane of contact.

Shield (щит). A continental block of the Earth's crust that has been relatively stable since Precambrian time; a disc-shaped volcano.

Shield volcano (щитовидний вулкан). A volcano composed of quiet lava flows.

Shoreline (берегова лінія, береговий контур). An average position of line where land and sea meet.

Siderophore (сидерофор). A nonporphyrin metabolite secreted by certain microorganisms and plant roots that forms a highly stable coordination compound with iron.

Silicate (силікат). Chemical compound which is a composition of silicon, oxygen, and some other element or elements.

Sill (сіл). A thin sheet of igneous rock intruded between and also parallel to older rock layers.

Silt (пил). 1) A soil separate consisting of particles between 0.05 and 0.002 millimeters in equivalent diameter. 2) A soil textural class.

Smectite (смектит, монтморилоніт). A group of 2:1 layer structured silicates with high CEC and variable interlayer spacing.

Smog (смог). A fog made heavier and darker by smoke and chemical fumes.

Sodic horizon (горизонт салік, соленосний шар). A mineral soil horizon of enrichment with secondary salts more soluble in cold water than gypsum.

Sodic soil (солонцевий ґрунт). A nonsaline soil containing sufficient exchangeable sodium to adversely affect crop production and soil structure.

Sodium adsorption (SAK). A relation between soluble cations, defined as:

$$SAR = \frac{(sodium)}{(calcium \oplus magnesium)^{\frac{1}{2}}}$$

and used to predict the exchangeable sodium percentage of soil equilibrated with a given solution. Concentrations, denoted by parentheses, are expressed in moles per liter.

Soil (ґрунт). 1) The unconsolidated mineral or material on the immediate surface of the earth that serves as a natural medium for the growth of land plants. 2) The unconsolidated mineral or organic matter on the surface of the earth, which has been subjected to and influenced by genetic and environmental factors of parent material, climate, macro- and microorganisms, and topography, all acting over a period of time and producing a product-soil-that differs from the material from which it is derived in many physical, chemical, biological, and morphological properties and characteristics.

Soil association (ґрунтова асоціація). A kind of map unit used in soil surveys comprised of delineation, each of which shows the size, shape, and location of a landscape unit composed of two or more kinds of component soils, or component soils and miscellaneous areas, plus allowable inclusions in either case.

Soil conservation (охорона ґрунтів). A combination of all management and land-use methods that safeguard the soil against depletion or deterioration by natural or human-induced factors.

Soil genesis (генезис ґрунтів). The made of origin of the soil with special reference to the processes of soil-forming factors responsible for the development of the solum, or true soil, from unconsolidated parent material.

Soil horizon (горизонт ґрунтового профілю). A layer of soil or soil material approximately parallel to the land surface and differing from adjacent genetically related layers in physical, chemical, and biological properties or characteristics such as color, structure, texture, consistency, kinds and number of organisms present, degree of acidity or alkalinity, and so on.

Soil loss tolerance (допустимі втрати ґрунту від ерозії). 1) The maximum average annual soil loss that will allow continuous cropping and maintain soil productivity without requiring additional management inputs. 2) The maximum soil erosion loss that is off set by the theoretical maximum rate of soil development, which will maintain an equilibrium between soil losses and gains.

Soil management groups (групи ґрунтів за спільними напрямками їх використання в сільському, лісовому господарстві та інших галузях економіки. Деякий аналог наших агровиробничих груп ґрунтів, але спектр використання тут ширший). Groups of taxonomic soil units with similar adaptations or management requirements for over or more specific purposes, such as adapted crops or crop rotations, drainage, fertilization, forestry, and highway engineering.

Soil monolith (ґрунтовий моноліт). A vertical section of a soil profile removed from the soil and mounted for display or study.

Soil productivity (продуктивність ґрунту). The capacity of a soil to produce a certain yield or crops, or other plants, with optimum management.

Soil profile (профіль грунту). A cross section of soil horizons.

Soil science (ґрунтознавство). That science dealing with soils as a natural resource on the surface of the earth, including soil formation, classification, mapping, geography and use, and physical, chemical, biological, and fertility properties of soils per se: and those properties in relation to their use and management.

Soil separates (гранулометричні фракції ґрунту). Mineral particles, less than 2.0 millimeters in equivalent diameter, ranging between specified size limits.

Soil series (ґрунтові серії американської класифікації ґрунтів). The lowest category in the U.S. system of soil taxonomy: a conceptualized class of soil bodies (polypedons) that have limits and ranges more restrictive than all higher taxa. The soil series serves as a major vehicle to transfer soil information and research knowledge from one soil area to another.

Soil solution (ґрунтовий розчин). The aqueous liquid phase of the soil and its solutes.

Soil structure (структура ґрунту). The combination or arrangement of primary soil particles into secondary particles, units, or peds.

S**oil survey** (ґрунтове обстеження). The systematic examination, description, classification, and mapping of soils in an area.

Soil texture (текстура (механічний, гранулометричний склад ґрунту). The relative proportion of the various soil separates in a soil.

Soil water potential (total) (повний потенціал ґрунтової вологи). The amount of work that must be done per unit quantity of pure water, at a specified elevation and at atmospheric pressure to transport reversible and isothermally an infinitesimal quantity of water from a pull of pure water to the soil water (at the point under consideration.

Solam (горизонти профілю ґрунту до материнської породи, взяті разом). The upper and most weathered part of the soil profile; the A, E, and B horizons.

Solar wind (сонячний вітер). A flow of ionized gas blown out of the Sun at high speed (20km/sec); mostly consisting of protons and electrons.

Solid (твердий). A state of matter which has a definite shape and volume because molecules cannot move freely from place to place.

Soluble (розчинний). Capable of being dissolved and carried in a solution.

Solution (розчин). Condition in which particles of a solid are dissolved in a liquid and cannot be separated by filtration.

Species (вид (біол.)). A class of individuals having common characteristics and capable of interbreeding.

Specific gravity (питома вага). A ratio of the mass of a body to the mass of an equal volume of water.

Spectroscope (спектроскоп). Instrument for examining the visible spectrum.

Spectrum (спектр). A sequence of electromagnetic waves ranging from gamma rays (smallest) to radio waves (largest). The visible spectrum is a very small part of the electromagnetic spectrum.

Spiral galaxy (спіральна галактика). Disc-shaped galaxy with arms rotating around the center and gradually receding from it.

Spit (тут коса, а не плювати). A sandbar connected to the shoreline (коса на зразок Тузли).

Spodic horizon (горизонт сподік, ілювіальний горизонт, що служить діагностичним). A mineral soil horizon that is characterized by the illuvial accumulation of amorphous materials composed of aluminum and organic carbon with or without iron.

Spring tide (весняний прилив). A high tide caused by position of Sun, Earth, and Moon in a straight line.

Stable rock (порода стійка до вивітрювання). A rock composed of minerals that resist chemical weathering.

Stack (стек). Vertical block of resistant rock cut off from the mainland by wave action.

Stalactite (сталактит). Deposit of calcium carbonate hanging from the roof or sides of a cave.

Stalagmite (сталагміт). Deposit of calcium carbonate built up from the floor of a cave by the drip of water from above.

Stratosphere (стратосфера). A layer of the Earth’s atmosphere from about 10km to 50km above the Earth’s surface having relatively constant temperature and little water content.

Streak (колір риски). A color of the fine powder of a mineral obtained by scratching it against a hard white surface.

Stress (стрес, тут: напруга у земній корі). The forces that push and pull on the Earth’s crust, causing its deformation.

Striation (штриховка від льодовика чи іншого твердого тіла). One of a group of parallel grooves caused by abrasion by ice or rock.

Strike-slip boundary (взаємодія між краями тектонічних плит, що рухаються один повз одного). A plate boundary at which two plates slip past one another horizontally.

Strip cropping (смугове вирощування культур). The practice of growing crops that require different types of tillage, such as row and sod, in alternate strips along contours or across the prevailing direction of the wind.

Subduction (пірнання під, залітання під, сабдукція). The process in which crust plunges back into interior of the Earth (занурення океанічної плити під материкову).

Subsoil (підґрунтя). The soil in the B horizon, or middle layer of soil.

Subsoiling (поглиблення орного чи іншого шару обробітку ґрунту). Any treatment to loosen soil below the tillage zone without inversion and with a minimum of mixing with the tilled zone.

Supernova (вибух зірки). A star explosion occurring near the end of a star's life.

Superposition (суперпозиція, хронологічний порядок залягання верств). An order in which rocks are deposited one above the other.

Surface charge density (густина заряду на поверхні мінералу). The excess of negative or positive charge per unit area of surface area of soil or soil mineral.

Suspension (суспензія). A state in which particles of a solid are mixed with a fluid but are undissolved.

S-wave (вторинні сейсмічні хвилі). Secondary earthquake wave which moves up and down perpendicular to the direction of wave travel.

Symmetrical (симетричний). Having parts that correspond in size, shape, and relative position on opposite sides of a dividing line or axis.

Syncline (синкліналь). A fold in a rock in which strata dip inward from both sides.

T

Tales (розсипище). A heap of rock debris at the floor of a cliff.

Tectonics (тектоніка). The branch of geology that deals with the movements that shape the Earth's crust.

Tension (напруга). A system of forces tending to pull a body apart; the stress resulting from such forces.

Terrace (тераса). A flat platform of rocks, sand, and silt at the base of a sea cliff (there are also the terraces of a river valley marking the former positions of the riverbed – the authors).

Tetragonal (тетрагональний). A crystal system having three axes at right angles to one another, two of equal length and one longer or shorter.

Tetrahedron (тетраедр). A solid bounded by four plane faces.

Texture (текстура породи; гранулометричний склад гранту у ґрунтознавстві). Characteristics of rock particles including size, shape, and arrangement of the particles.

Theory (теорія). A principle based on a body of facts which has stood the test of time.

Theory of continental drift (теорія континентального дрейфу). The theory, proposed by Alfred Wegener, that the continents were once joined together and have since drifted apart.

Theory of platetectonics (теорія тектоніки плит). The theory that links together the ideas of continental drift and ocean-floor spreading and explains how the Earth has changed over time.

Thermal (термальний). Of or pertaining to heat.

Thermic (термік, тип теплового режиму ґрунту). A soil temperature regime that has mean annual soil temperatures of 15°C or more, but less than 22°C and more than 5°C difference between mean summer and mean winter soil temperatures at a 50-centimeter depth below the surface.

Thermosphere (термосфера). A layer of the atmosphere between the mesosphere and the exosphere.

Thrust fault (розривна дислокація). A reverse fault in which the hanging wall slides over the foot wall.

Tidal (що має відношення до приливів). Of or pertaining to the periodic rise and fall of waters of the ocean.

Till (моренний матеріал, льодовикові відклади). 1). Unstratified glacial drift deposited by ice and consisting of clay, silt, sand, gravel, and bounders, intermingled in any proportion. 2). To prepare the soil for seeding; to seed or cultivate the soil.

Tilth (складення ґрунту, агрономічно сприятливий стан щільності та пористості ґрунту). The physical condition of soil as related to its ease of tillage, fitness as a seedbed, and its impendence to seedling emergence and root penetration.

Top dressing (поверхневе підживлення (добривами)). An application of fertilizer to a soil surface, without incorporation, after the crop stand has been established.

Topography (тут: рельєф). Surface features of an area.

Toposequence (катена, послідовність одиниць ґрунтового покриву за рельєфом). A sequence of related soils. The soils differ, one from the other, primarily because of topography as a soil-forming factor.

Topsoil (верхній шар гранту). The soil in the A horizon, or uppermost layer of mature soil.

Tornado (торнадо). A violent storm formed when a mass of warm air is trapped between layers of cold air and air rushes up in a spiral pattern.

Torric (тип водного режиму ґрунту). A soil-moisture regime defined like aridic moisture regime but used in a different category of the U.S. taxonomy.

Transform fault (дислокація на дні океану). A fault that runs across a mid-ocean ridge.

Transported soil. Soil that is moved away from its place of origin.

Trellis (форма річкової мережі). Stream pattern of tributaries parallel to each other.

Trench (жолоб океанічного дна). A V-shaped valley on the ocean floor where old ocean floor is subducted; a destructive (convergent) plate boundary.

Tributary (притока). (1) a large stream or small river that flows into an area’s main river. (2) A stream contributing water to another larger stream.

Tropics (тропіки). A zone of the Earth’s surface extending 23-1/2° north and 23-1/2° south on either side of the equator.

Tropopause (тропопауза). A layer of the Earth’s atmosphere between the troposphere and the stratosphere where air is calm and storm-free.

Troposphere (тропосфера). A layer of the Earth’s atmosphere from the surface to about 10 km above the Earth containing about 75 % of the gases of the atmosphere.

Truncated (вкорочений, зрізаний (профіль ґрунту)). Having lost all or part of the upper soil horizon or horizons.

Tsunami [su’nʌ:mɪ] (цунамі). A great sea wave produced by a submarine earthquake.

Tuff (туф). Volcanic ash usually more or less stratified and in various states of consolidation.

Turbidity current (каламутна океанічна течія, спричинена різною щільністю води, що містить каламуть). An ocean current caused by density of water containing sediment in suspension.

U

Udic (тип водного режиму ґрунту). A soil moisture regime that is neither dry for as long as 90 cumulative days nor for as long as 60 consecutive days in the 90 days following the summer solstice (сонцестояння) at periods when soil temperature at 50 centimeters below the surface is above 5°C.

Ultisols (ультисолі, ґрунтовий порядок). Mineral soils that have an argillic or kandic horizon with a basic cation saturation of less than 35 percent when measured at pH 8.2. Ultisols have a man annual temperature of 8°C or higher.

Umbra. A part of a shadow in which the light from the source is completely blocked out.

Umbric epipedon (поверхневий горизонт умбрік). A surface layer of mineral soil that has the same requirements as the mollic epipedon with respect to color, thickness, organic carbon content, consistence, structure, and phosphorus content, at that has a basic cation saturation less than 50 percent when measured at pH 7.

Unconformity (невідповідність). A break in the rock record.

Universe (всесвіт). All bodies in the space and all the space in between those bodies.

Upwelling (переміщення води до гори). A process by which deep, cold water is brought to the surface.

Ustic (тип водного режиму ґрунту). A soil moisture regime that is intermediate between the aridic and udic regimes and common in temperature subhumid or semiarid regions, or in tropical and subtropical regions with a monsoon climate. A limited amount of water is available for plants but occurs at times when the soil temperature is optimum for plant growth.

V

Vein (жила, прожилок). A narrow, well defined zone containing mineral-bearing rock in place.

Velocity (швидкість). A time rate of linear motion in a given direction.

Vent (отвір у кратері). An opening through which lava erupts.

Vermiculite (вермикуліт, мінерал). A highly charged layer-structured silicate of the 2:1 type that is formed from mica.

Vertegrate (хребетний). An animal having a spinal column.

Viscous [ʻvɪskəs] (в'язкий, густий (про флюїд)). Pertaining to a fluid which resists flowing because of its thickness.

Volcanic ash (вулканічний попіл). Rock particles more than 0.25mm but less than 5mm across that are blown into the air by a volcanic eruption.

Volcano (вулкан). A place in the Earth's surface through which molten rock and other materials reach the surface.

Volume (об'єм). The space that an object occupies or fills; expressed in cubic units such as cubic centimeters (cm^3) or milliliters (mL).

W

Water cycle (коло біг води). A continuous pattern of water circulation from sea to atmosphere, to land. And to sea again.

Water potential (водний потенціал, див. Soil water potential).

Water table (рівень ґрунтових вод). The upper surface of groundwater or that level in the ground where the water is at atmospheric pressure.

Water table, perched (рівень ґрунтової вадозної води, верховодки). The water table of a saturated layer of soil that is separated from an underlying saturated layer by an unsaturated layer (vadose water).

Waterlogged (підтоплений знизу (ґрунт)). Saturated or nearly saturated with water.

Wavelength (довжина хвилі). A horizontal distance between similar points on two successive waves measured perpendicularly to the crest.

Weather (погода). Day-today changes in wind, temperature, humidity, and pressure.

Weathering (вивітрювання). A physical disintegration and chemical decomposition of earth material at or near the surface.

Weight (вага). Measure of gravitational pull of the Earth on bodies near the surface of the Earth.

Weightlessness (невагомість). A state of an object when a force acting on it is equal and opposite to the Earth's gravity or when it is outside of the Earth's gravity or the gravity of any other large mass.

Wilting point (вологість в'янення). See permanent wilting point (PWP).

Wind chill (вітерцева прохолода, зниження температури від вітру). A still-air temperature that would have the same cooling effect on exposed human flesh as a given combination of temperature and wind speed (it is important for temperatures well below freezing point).

Windward (напрям (бік), з якого дує вітер; завітрений бік, наприклад, схилу). A direction from which the wind is blowing.

X

Xeric (тип водного режиму ґрунту). A soil moisture regime common to Mediterranean climates having moist, cool winters and warm, dry summers. A limited amount of water is present but does not occur at optimum periods for plant growth. Irrigation of summer fallow is commonly necessary for crop production.

Xerophytes (ксерофіти). Plants that grow in or on extremely dry soils.

X-rays (рентгенівські промені). Electromagnetic radiation of very short wavelength and high energy.

Y

Yield (врожайність). The amount of a specified substance produced (e.g., grain, straw, total dry matter) per unit area.

Z

Zero point of charge (ізотонічне значення pH у розчині). The pH value of a solution in equilibrium with a particle whose net charge from all sources is zero.

Zero tillage (нульовий обробіток ґрунту). See No-tillage system.

Zone of saturation (зона насичення). A zone where all rock openings are filled with fluid.

Appendix One

Made Simple Variant of Plate Tectonics Theory

(from FOCUS ON EARTH SCIENCE, Edited by A.Cameron. published by C.E. Merril Publishing Co. Columbus, Ohio 43216, 1981, p. 303-321)

Continents have three major features. The oldest area of each continent is the **shield**. This is an area of low lands. Surrounding the shield is the **stable platform**. This area is covered by relatively undisturbed sedimentary rocks. Along the continental margins are **folded mountains**. All continents have folded mountains on at least one coast. Here sedimentary rocks are complexly folded, fractured, and faulted.

Rocks of folded mountains are relatively young compared to the continent's nucleus. They are mostly shallow water sedimentary materials, but they have great thicknesses. In the mountain core, sedimentary rock may be intruded by igneous rocks, or metamorphosed. The great bolds of long mountain chains are evidence of movement of the earth crust. Movements of the crust is slow, but it can be measured in centimeters per year.

Ocean floors are rugged surfaces. A long system of ridges, circling the globe, were discovered. The ridge system extends from Iceland in the Arctic, through the Atlantic and Indian Oceans into the Pacific. It crosses the Pacific and parallels the coast of South America until it reaches the Gulf of California. The Gulf of California is new ocean. This segment of the **mid-ocean ridge** is joined to another segment by the San Andreas fault. The second segment begins off the coast of Northern California and runs northward to Alaska.

The other major features of the ocean bottom are the deep sea trenches. Most of them are close to continental borders. Geologists for many years have known that uplifted mountains contain thousands of meters of shallow water sediments. Typically, limestones, sandstones, and shales containing fossils are deposited in the near shore zone. Yet folded mountains have great thicknesses of such layers. The conclusion is that the shallow water sediments were deposited in a sinking trough near the shore zone. At some depth, equilibrium was upset. The less dense materials rose to form mountains.

Plate tectonics is the newest explanation for changes in the earth's crust caused by forces from within the earth. In 1911 Alfred Wegener, a German meteorologist, noticed that Africa and South America appeared to fit together like pieces of a giant puzzle (jigsaw puzzle = a picture stuck onto wood and cut up into many small irregular pieces to be fitted together for amusement). Frank Taylor, an American geologist, also wrote about his ideas of drifting continents. Wegener suggested that Africa and South America had been joined at some time. Then they broke apart and drifted to their present positions. Evidence from biology also supported the idea of drifting continents. Glossopteris (and extinct plant) fossils, which are located in rocks about 250 million years old, are found in South Africa, Australia, India, and Antarctica. The seeds were too large to have been carried by wind and too fragile to have survived a trip by ocean waves. The places in which the plant's fossils are found must once have been closer together. There is also an evidence from rocks. And the evidence of the ocean floor spreading: as the ocean floor moves away from the midocean ridge, lava flows out of the rift and hardens to form new ocean floor.

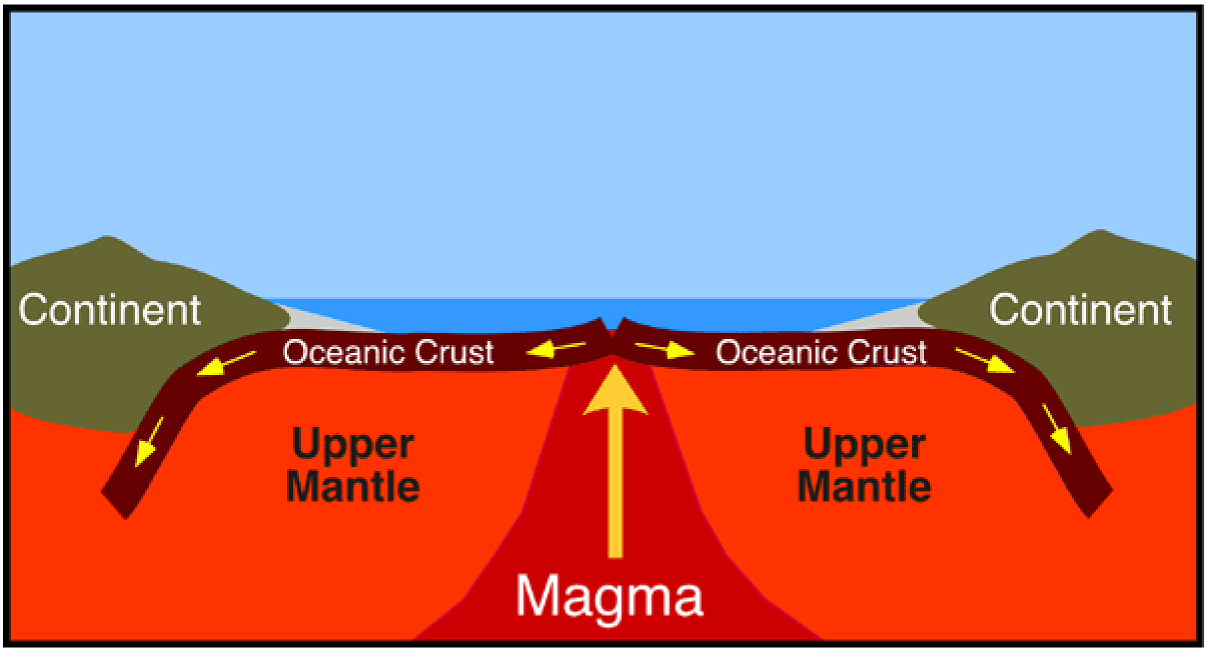


Fig. 99. Ocean Floor Spreading (source: www.mlhi.org/science/period6/Sea.htm).

New deep-sea drilling machines also provided evidence to support the idea of ocean floor spreading. Rock samples from the ocean floor indicate that rocks next to a midocean ridge are younger than rocks farther away. The youngest rocks are in the center of the ridge. As the ocean floor spreads, the older rocks move farther away from the ridge.

You might think that as a result of ocean-floor spreading, the Earth's surface is getting larger. Because the Earth's surface remains the same size, the ocean floor is being destroyed as fast as it I being formed by ocean floor has been destroyed. But how does this destruction occur? The answer involves deep, V-shaped valleys called trenches that lie along the bottom of the oceans. Older ocean floor moves away from the midocean ridges as new ocean floor is formed. Eventually, the older ocean floor moves down deep into the Earth along the trenches. The process, in which crust plunges back into the Earth, is called subduction. As new rocks are being formed along the midocean ridges, older rocks are subducted into the trenches. One process balances the other. The Earth's crust remains the same size (Fig. 100).

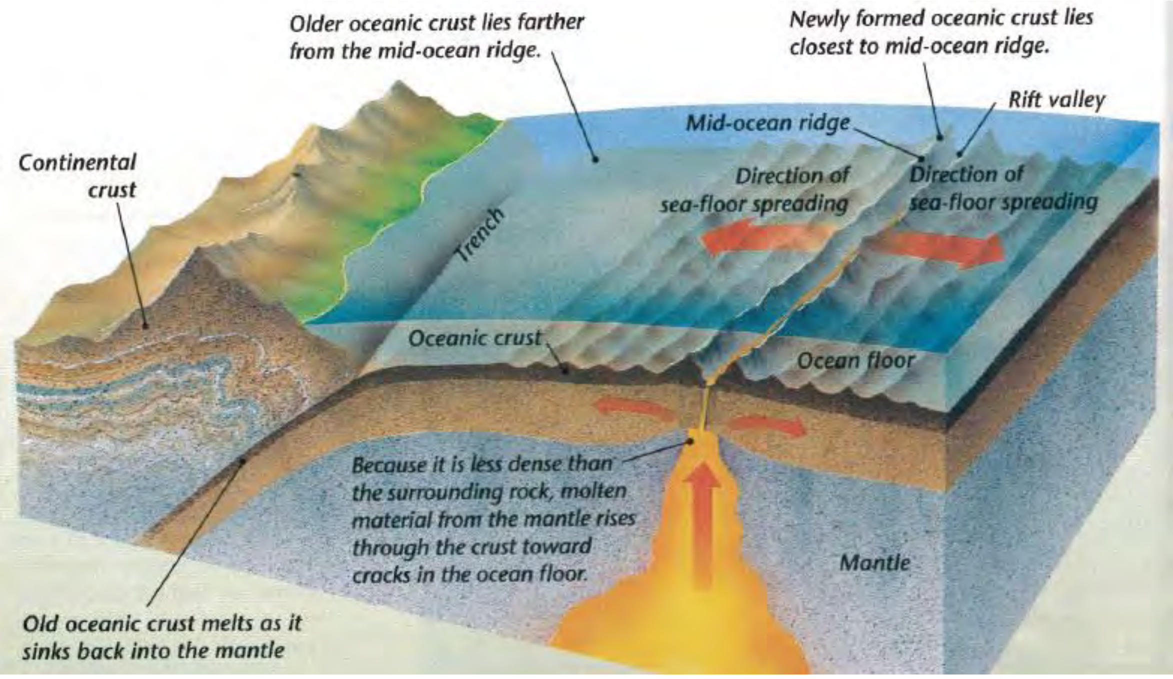


Fig. 100. Creation and Destruction of the Ocean Floor.

The overwhelming evidence of these phenomena caused many of the old theories to be discarded. The theory of plate tectonics appeared. The Earth's crust is made of huge floating plates. The plates carry the continents and are edged by trenches and ridges. The theory of plate tectonics, which links together the ideas of continental drift and ocean-floor spreading, explains how plain the formation, movement, collisions, and destruction of the Earth's crust.

The theory provides a framework for understanding mountains, volcanoes, earthquakes, and other landforms and processes of the physical Earth.

There are seven major lithospheric plates, each of which is named after its surface features. The Pacific plate, which covers one-fifth of the Earth's surface, is the largest plate. The other major plates are the North American, South American, Eurasian, African, Indo-Australian, and Antarctic plates. There are also many smaller plates. Some of these, such as the Caribbean and Arabian plates, are fairly large.

Others are so small that they are not included in maps that show the entire Earth. Plates move at different speeds and in different directions. Some move only a few millimeters per year. In a few cases, the edges of the continents are the boundaries of plates. Most plate boundaries are on the ocean floor.

There are three types of plate boundaries. The first type occurs at Midocean ridges. Because plates move apart (diverse) at Midocean ridges, the ridges are called divergent boundaries. These boundaries are also called constructive boundaries. Because new ocean floor forms. The second type of plate boundary has trenches. The plates come together (converge) at the trenches. The trenches are called convergent boundaries. They are also called destructive boundaries because old ocean floor is destroyed. The collision of plates at such boundaries causes tremendous pressure and friction. Severe earthquakes often result. As plate material melts in the Earth's mantle, some of it surges upward to produce volcanoes.

The third type of plate boundary is formed by a lateral fault. Boundaries formed by lateral faults are called strike-slip boundaries. Two plates grind together and slip past each other horizontally. No new plate material is made, and no plate material is destroyed. Such a boundary is called a conservative boundary. Earthquakes often occur along strike-slip boundaries. Scientists are not sure exactly what makes the plates move. One hypotheses is that large convection currents within the Earth move the plates.

Plate tectonics has a connection to the development of life on Earth. Living things evolve in response to changes in their environment. And the movement of the plates causes changes in climate, in geographic features such as mountains and in the types of living things with which a species interacts. When landmasses join together, diversity decreases. For example, fossils indicate that there were once 29 families of mammals in South America and 27 entirely different families of mammals in North America. Soon after the continents joined together – about 3 million years ago – there were only 22 families left. Only the families that competed the most successfully survived. The rest died out. When landmasses split apart, the diversity of land animals increases. On a big landmass, animals can easily move to suitable places and avoid the more challenging environments. On a small landmass, animals are stuck where they are and thus must adapt to local conditions. The splitting up of South America and Africa roughly 45 million years ago resulted in monkeys evolving into two distinct groups. New world monkeys are primarily tree-dwellers that have long tails used for grasping and for balance. Although old world monkeys include tree-dwellers as well as ground dwellers, none has a grasping tail.

Appendix Two

FAO and WRB Nomenclature Names of the Most Widespread Mapping Units of Ukrainian Soils (From Medvedev, et el., 2003)

Ukrainian Name	Word for Word Translocation into English	Soil Name in WRB (1998) Nomenclature	Soil Name in FAO (1997) Nomenclature
Дерново-слабопідзолисті піщані та глинисто-піщані	Soddy slightly podzolic soils in loose and bound sand deposits	Albeluvisols Umbric (ABum)	Eutric Podzoluvisols (Pde)
Дерново-середньопідзолисті супіщані	Soddy moderately podzolic soils in loamy sand deposits	>> – >>	>> – >>
Ясно-сірі та сірі лісові ґрунти	Light grey and grey forest soils	Albeluvisols Umbric (ABum), Phaozems Albic (Phab)	Eutric Podzoluvisols, Haplic Greyzems (Pde, GRh)
Темно-сірі опідзолені ґрунти	Dark grey podzolized soils	Phaozems Albic (Phab)	Haplic Greyzems (GRh)
Чорноземи опідзолені	Podzolized chernozems	Chernozems chermic (CHch)	Haplic chernozems (CHh)
Чорноземи типові на лесах та лесовидних суглинках	Typical Chernozems in loess and loess-like loams	>> – >>	>> – >>
Чорноземи звичайні на лесах та лесоподібних породах	Ordinary Chernozems in loess and loess-like parent materials	>> – >>	>> – >>

Чорноземи південні на лесах та лесоподібних породах	Southern Chernozems in loess and loess-like parent materials	Chernozems Calcic (CHcc)	Calcid Chernozems (CHc)
Темно-каштанові залишково-солонцюваті ґрунти	Dark Chesnut Soils with residual sodicity	Kastanozems Haplic (Ksha)	Haplic Kastanozems (KSh)
Каштанові солонцюваті ґрунти	Chesnut sodicity-affected soils	Kastanozems Luvic (KSlv)	Luvic Kastanozems (KS l)
Лучні ґрунти на алювіальних відкладах	Meadow soils in alluvial deposits	Umbisols Gleic, Fluvisols Dystric (HMgl, FL dy)	Umbric Gleysols Dystric Fluvisols (Gle, FLd)
Торфово-болотні грунти та торфовища	Peat-Bog and Peat Soils	Histosols Sapric (HS sa)	Terric Histosols (HSs) Histosols Histosols (HS)
Солонці	Alkali soils, Sodic Soils	Solonetz Humic (SN hu)	Haplic Solonetz (SNh)
Осолоділі грунти	Solodized Soils	Planosols Albic (PL ab)	Eutric Planosols (PLe)
Дернові грунти	Soddy Soils	Arenosols Haplic (AR ha)	Haplic Arenosols (Arh)
Бурі гірсько-лісові грунти	Brown mountain forest soils	Cambisols Dystric (CM dy)	Dystric Cambisols (CMd)
Коричневі гірські грунти на елювії – делювії корінних порід	Cinnamonic mountain soils in eluvium-deluvium of consolidated rocks	Cambisols Calcaric (CM ca)	Calcaric Cambisols (CMc)

Appendix Three

Ecologico-Substantive Soil Classification in Ukraine (M.I.Polupan, et al., 2005)

Type	Subtype	Genus	Species	Variant	Lithologic Series
1. Soddy-Podzolic	–	Light Sand to Medium Loam	Automorphic, slightly gleyed, gleyed	Virgin, reclaimed, modal, cultured, eroded, drained, wet as a result of irrigation	Glacio-fluvial, morainic, loess, ancient alluvial deposits
2. Soddy Podzolized	–	Sand to heavy medium loam	The same, carbonatic, skeletal	The same	The same, chalky-marl
3. Soddy Gleyed	–	The same	Gleyed, slightly gleyed, severely gleyed, carbonatic, shallow, short, non-deep	Virgin, turned to cultivation, modal, drained	The same
4. Light Grey Forest	Moderately slightly humuso-accumulative, low humuso-accumulative, slightly humuso-accumulative, very low humuso-accumulative,	Heavy loamy sand to heavy medium loam	Automorphic, gleyed from the surface, with excessive wetness, slightly xeromorphic, moderately xeromorphic, and very xeromorphic	Virgin, turned to cultivation: modal, eroded, drained, cultured, and deeply plowed	Loess, loess-like loam, ancient alluvial deposits

Type	Subtype	Genus	Species	Variant	Lithologic Series
5. Grey Forest	Moderately humuso-accumulative, moderately slightly humuso-accumulative, slightly humuso-accumulative, low humuso-accumulative,	Heavy loamy sand to light heavy loam	The same without gleyed from the surface	The same	The same
6. Dark Grey Podzolized	Moderately well humuso-accumulative, averagely humuso-accumulative, moderately slightly humuso-accumulative, slightly humuso-accumulative,	Sandy light loam to light clay	Slightly gleyed from the surface, automorphic, excessively wet, slightly xeromorphic, averagely xeromorphic, very xeromorphic	Virgin, put to cultivation: modal, deeply plowed, mocharic, mocharic-slightly gleyed, and mocharic-gleyed	Loess, loess-like loam
7. Podzolized Chernozem	Well humuso-accumulative, moderately well humuso-accumulative, moderately (averagely) humuso-accumulative, moderately slightly humuso-accumulative	The same	The same	The same	The same

Type	Subtype	Genus	Species	Variant	Lithologic Series
8. Typical Chernozem	Moderately highly humuso-accumulative, very well humuso-accumulative, well humuso-accumulative	Heavy loamy sand to light clay	Deep, excessively deep, overdeep, modal, excessively moist, by the extent of xeromorphicity	Virgin, reclaimed: modal, irrigated, eroded, with irrigation – caused sodicity, with irrigation – caused wetness, mocharik (with water seepage), mocharic– gleyous, mocharic - gleic	Loess, loess-like, ancient alluvial deposits
9. Meadow-typically-chernozemic	Highly humuso-accumulative, moderately highly	Light loamy sand – light clay	Excessively deep, overdeep, deeply solodized, by the extent of sodicity and salinity	Virgin, reclaimed, modal, irrigated, drained, salinated, and sodicity – affected as a result of irrigation	Loess, loess- like ancient alluvial, and clayey deposits

Type	Subtype	Genus	Species	Variant	Lithologic Series
10. Ordinary Chernozem	Moderately well humuso-accumulative, averagely humuso-accumulative, moderately weakly humuso-accumulative	Heavy loamy sand to medium clay	Overdeep, moderately deep, averagely deep, deep, modal, excessively moist, by the extent of xeromorphicity and stonyness	Virgin, cultivated, modal, irrigated, eroded, irrigation caused sodicity affected, wet as a result of irrigation, deeply plowed, mocharic (with water seepage), mocharic-gleyous, and mocharic-gleyic	Loess, loess-like, ancient alluvial deposits, clay, sand-stony, clayey-slatic, chalk-marlic
11. Meadow-ordinary chernozemic	Well humuso-accumulative, moderately well humuso-accumulative, averagely humuso-accumulative	Light loamy sand to light medium clay	Moderately deep, deep, excessively deep, by the extent of sodicity and salinity	Virgin, cultivated, modal, irrigated, sodicity or salinity – affected as a result of irrigation, drained	Loess, loess-like, ancient alluvial
12. Southern Chernozem	Weakly humuso-accumulative	Heavy loamy sand to light clay	Nondeep, averagely deep, moderately deep, modal, excessively moist, by the extent of xeromorphicity and skeletality (stonyness)	Virgin, cultivated, modal, irrigated, gleyed from the surface, as a result of irrigation, affected by irrigation – caused sodicity, wet as a result of irrigation, deeply plowed	Loess, ancient alluvial, chalk-marlic, sandstony

Type	Subtype	Genus	Species	Variant	Lithologic Series
13. Meadow-Southern Chernozemic	Moderately Slightly humuso-accumulative	Heavy loamy sand to medium clay	Moderately deep, averagely deep; by the extent of sodicity and solinity	Virgin, cultivated, modal, irrigated, sodicity affected and saline as a result of irrigation	Loess, ancient alluvial deposits, clay, chalky-marl, limestone
14. Chernozemic Meadow	Highly humuso-accumulative, moderately highly humuso-accumulative, very good humuso-accumulative, well humuso-accumulative, moderately humuso-accumulative, moderately slightly –humuso-accumulative	The same	Modal, sodicity affected, solonchaky (very saline), sodic, solod-like	The same	Loess, ancient alluvial deposits
15. Meadow Southern Chernozemic gleyed from the surface	Slightly humuso-accumulative	Heavy medium loam to light clay	Averagely deep, gleyous, gleyed, carbonatic, slightly solodized, moderately solodized	Virgin, cultivated, modal, wet as a result of irrigation	Gleyed loess
16. Soddy gleyed from the surface	Very slightly humuso-accumulative	Heavy clay	Gleyed, modal, carbonatic, slightly solodized	Virgin, reclaimed, modal, waterlogged as a result of irrigation	The same

Type	Subtype	Genus	Species	Variant	Lithologic Series
17. Soddy gleyed from the surface and solodized	Very slightly humuso-accumulative	Light heavy loam to medium clay	Gleyed; by the extent solodizity: slightly, moderately, severely, and very severely solodized	The same	The same
18. Dark chestnut soil	Slightly humuso-accumulative	Heavy loamy sand to medium clay	Not deep, averagely deep, modal, excessively moist, by the extent of xeromorphicity; slightly, moderately, and very xeromorphic	Virgin, reclaimed; modal, eroded, irrigated, sodicity-affected from the surface and may be salinated as a result of irrigation, wet from irrigation	Loess, loess-like loam, clay, alluvial
19. Sodicity-affected chestnut soil	Very slightly humuso-accumulative	Light loamy-sand to medium clay	Not deep, averagely deep, modal, excessively moist, by the extent of xeromorphicity	Virgin, cultivated, modal, eroded, irrigated, sodicity-affected by irrigation, saline and sodicity-affected by irrigation, wet as a result of irrigation, gleyed from the surface as a result of irrigation, deeply plowed	Loess, loess-like loam, clay, ancient alluvial deposits

Type	Subtype	Genus	Species	Variant	Lithologic Series
20. Chestnut Sodic soil	With ultra-low humuso-accumulation	Light loam to light clay	By the depth of eluviated horizon: shallow, moderately deep, and deep	The same	The same
21. Meadow-chestnut sodicity-affected	Very slightly humuso-accumulative	Light heavy loam to medium clay	Not deep, with excessive wetness, sodicity and salinity-affected	The same	Loess, marine deposits
22. Sodic meadow chernozemic	Ultra low humuso-accumulative	The same	Moderate, deep, salinity-affected from certain depth or from the very surface	The same	The same
23. Chestnut meadow sodicity and salinity-affected	With very low humuso-accumulation	Heavy medium loam to light clay	Not deep, saline from certain depth or from the surface	Virgin, cultivated, modal, gleyed from the surface as a result of irrigation, drained, deeply plowed	Loess
24. Chestnut saline-sodic	Ultralow humus accumulation	The same	Shallow, moderately deep, gleyed and solodized at the same time	The same	The same
25. Soddy gleyed saline	With very low humus accumulation	Light to heavy clay	Not deep, short, modal, solodized	Virgin	Gleyed loess
26. Very saline gleyed	Ultralow humus accumulation	The same	Modal	Virgin	The same

Type	Subtype	Genus	Species	Variant	Lithologic Series
27. Brown soil	Extrahumuso-accumulative or subextrahumuso-accumulative	Sandy light loam to light medium loam	Shallow, short, not deep, moderately deep, by the extent of “skeletality”	Virgin, virgin but cultured	Sandstone, slate, gneiss, etc.
28. Podzolized brown soil	Very high humus accumulation, high humus accumulation, very good humus accumulation, moderately good humus accumulation	Sandy light loam to heavy loam	Shallow, short, not deep, moderately deep, averagely deep, automorphic, slightly gleyed, acid, slightly saturated, carbonatic, by the extent of “skeletality”	Virgin, reclaimed, modal, cultured	Sandstone, clay-slate, limestone, volcanic tufa, argillite, andesitic – basalt, gneiss, etc
29. Brown Podzolic gleyed from the surface	Low humus accumulation	Light medium loam – light clay	Averagely deep, moderately deep, acid, slightly saturated, by the presence of skeletal particles	Virgin, reclaimed, modal, cultured, drained	Clay and slate, andesite and basalt, volcanic tufa, loess-like loam, diluvium, etc.
30. Meadow brown podzolized, gleyed from the surface	Moderately slightly humuso-accumulative	Light medium medium loam to light heavy clay	Deep, more than averagely deep	Virgin, reclaimed, cultured, modal, drained	Alluvial, lacustrine-aluvial

Type	Subtype	Genus	Species	Variant	Lithologic Series
31. Meadow brown gleic	The same	Heavy loam to light clay	Gleyed	The same	The same
32. Cinnamon	Slight or low humus accumulation	Light loam to medium clay	Not deep, averagely deep, by the extent of salinity, sodicity and stonyness	Virgin, reclaimed, modal, eroded, deeply plowed	Limestone, slate, conglomerate, massive crystalline, etc.
33. Meadow swampy	–	Light loamy sand to light clay	Short, not deep, carbonatic, sodicity affected and saline from certain depth	Virgin, reclaimed, modal, drained, saline as a result irrigation	Loess, loess-like loams, clay, ancient alluvium
34. Swamp (muck)	–	The same	Not deep, moderately deep, carbonatic, saline sodicity-affected	Virgin, reclaimed, dried, drained	The same
35. Alluvial meadow	Accumulative, podzolized, brown soil – podzolized	Sand to light clay	Short, not deep, averagely deep, layered, by the extent of sodicity, salinity and stonyness	Virgin, reclaimed, modal, drained, irrigated, sodicity affected as a result of irrigation	Alluvial deposits, pebbles, marl

Type	Subtype	Genus	Species	Variant	Lithologic Series
36. Alluvial meadow-swampy	–	The same	Short, not deep, layered, salinity and sodicity affected	Virgin, reclaimed, drained, cultured	Alluvial deposits, marl
37. Alluvial swampy	–	Light loam to light clay	Short, not deep, moderately deep, layered, carbonatic sodicity and salinity affected	The same	The same
38. Lowland Peat	Shallow peaty gleyed, peaty gleyed, and peaty	Low ash content, moderate ash content, and high ash content	Not deep, moderately deep, averagely deep, deep, more than averagely deep, excessively deep, shallowly buried, by the extent of peat decomposition, mineral-gleic, carbonatic, ironous, sodicity and salinity affected	Virgin, reclaimed, drained, humisified, mineralized	Alluvial deposits, marl, ancient alluvium

Type	Subtype	Genus	Species	Variant	Lithologic Series
39. Recultizem	Pollisian sufficiently moist, forest-steppe moist, very moist, sufficiently moist, excessively moist, moderately moist, etc.; steppe insufficiently moist, moderately draughty, semiarid, arid, southern steppe moderately draughty; arid steppe arid and very arid; brown-soil very wet, moderately wet, and normally wet, xerophytic forest	Fine sand to medium clay	Humus-provided, shallow, moderately deep, petromorphic (without humus layer); by the extent of stonyness and gleying	Modal, irrigated, polluted	Loess, loess-like loam, ancient alluvium, moraine, glacial till, clay and slate, chalk and marl, etc.
40. Mocharic chernozemic meadow	Are intensified in zonal and subzonal aspects in accordance with “background” soil	Heavy loamy sand to medium clay	By the depth of profile and the extent of soil sodicity and salinity	Virgin, reclaimed, modal, and drained	Loess, clay
41. Mocharic meadow swampy	The same	The same	The same	The same	The same
42. Mocharic swampy	The same	The same	The same	The same	The same
43. Soddy mocharic-gleic	The same	Heavy loam to medium clay	By the depth of soil profile	The same	The same

Appendix Four

Ukrainian Soils and Landscapes

Forest (Polissya) Zone

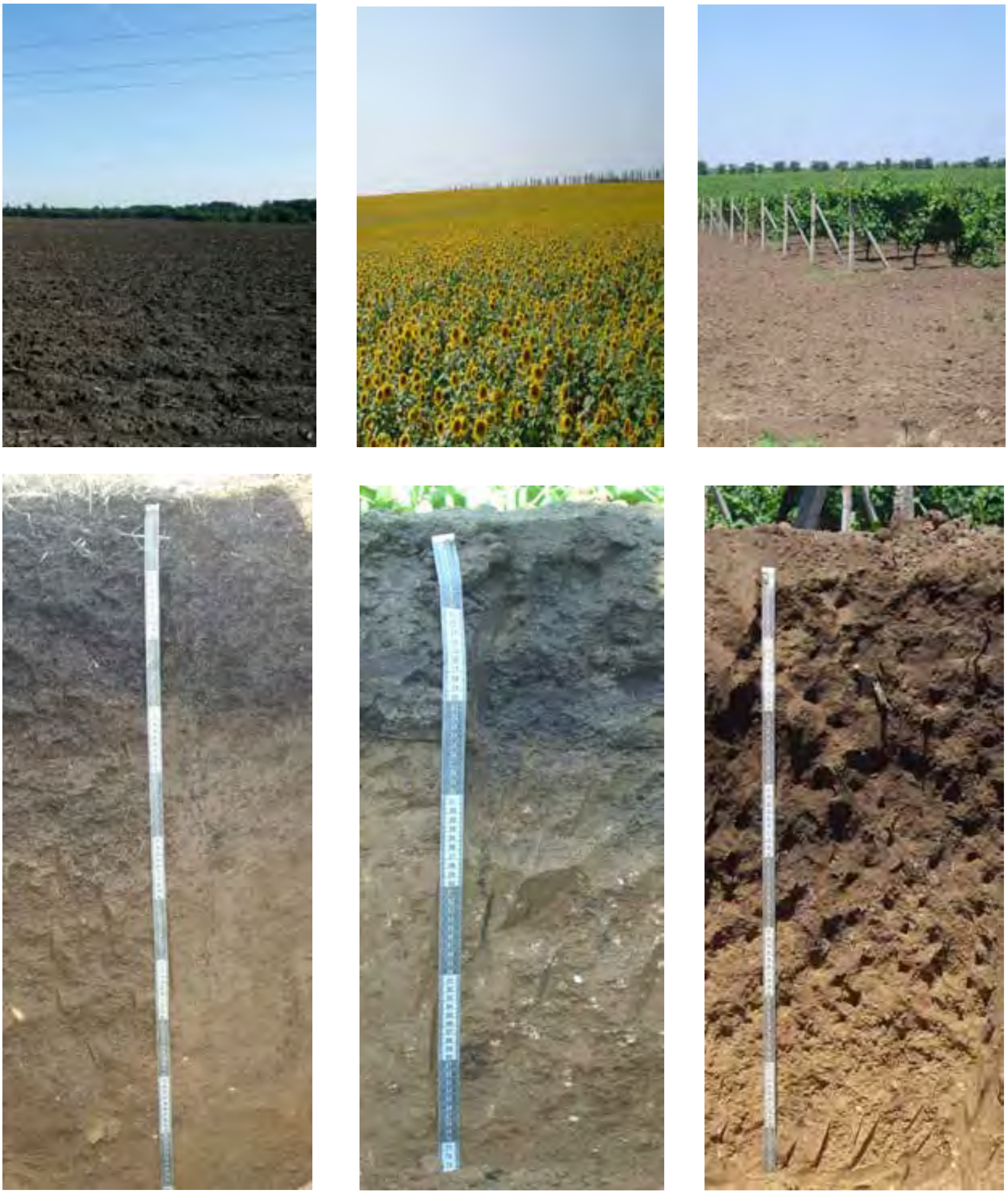
Sod-Podzolic

Podzolic-Sod

Peat

Forest-Steppe Zone

Light Grey **Podzolized Chernozem** **Typical Chernozem**



Steppe Zone

Ordinary Chernozem **Southern Chernozem** **Dark Chestnuts Soil**

Mounting Zone

Very Saline (Solonchak) Soil

Red-Brown Soil of Xerophytic Forests

Carpathian Brown Forest Soil